

SEISMIC STUDY OF THE CRUST AND UPPER MANTLE STRUCTURE IN EASTERN
ASIA USING AMBIENT NOISE CORRELATION AND EARTHQUAKE DATA

BY

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DISSERTATION

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ABSTRACT

Seismic methods using natural earthquakes are commonly applied to investigate the velocity structures of the crust and upper mantle. Among all these methods, seismic tomography is one of the most powerful tools to image the earth's structure. Recent theoretical and experimental studies suggest that the media impulse response (Green's function) of surface wave between two seismic stations can be retrieved from the cross correlation of diffusive wave field. This method provides additional data coverage to the study region and has been quickly used in various seismic studies. In this dissertation, I develop several tomography methods using traditional earthquake data as well as surface wave data from ambient noise correlation and apply them in eastern Asia.

A joint inversion scheme using first and secondary Pg arrivals and Pn arrivals is developed to determine the crustal and uppermost mantle P wave velocity and the depth of Moho discontinuity. This method addresses the trade-off issue between the media velocity and a discontinuity location. I apply this method to the eastern and southern margin of Tibetan Plateau, where the Moho depth changes rapidly. Significant features are revealed from tomography results, in consistence of regional geological settings, including slow upper crust but fast lower crust under Sichuan Basin, rapid Moho variation, and slow Pn in the plateau and fast Pn in the borderland separated by major boundaries and faults. The mid-lower crust in the plateau is relatively slow compared to the average model and Sichuan Basin and Poisson's ratio is anomalous (exceeding 0.3), which is consistent with weak mid-lower crust in the plateau.

The crust and upper mantle shear wave velocity in China and its surrounding region is obtained from regional surface wave tomography. By combining traditional earthquake surface wave data with ambient noise correlation data, I am able to extend the Rayleigh wave group velocity period from 8 s to 120 s. Together with Rayleigh phase velocity from 8 s to 70 s derived from ambient noise data, shear velocity down to 150 km is inverted in the study area. Striking crust and upper mantle structures are well resolved, including major basins, Moho depth variation, mantle velocity contrast between eastern and western North China craton, widespread low-velocity zone in mid-crust in much of the Tibetan Plateau, and clear velocity contrasts of the mantle lithosphere between north and southern Tibet with significant E-W variations.

To overcome the depth-velocity trade-off of teleseismic P wave receiver function and the insensitivity to velocity contrast in surface wave inversion, a nonlinear searching scheme using Neighborhood Algorithm is developed to jointly determine shear velocity and Moho depth using these two very different data sets. This method is applied to Hi-CLIMB seismic stations in Tibetan Plateau. A high resolution 2D shear velocity image is derived from joint inversion. The results reveal the detailed shape of mid crust low velocity zone that exists underneath most part of Tibet and the Moho variations between Lhasa block and Qiangtang block. A zone of interrupted Moho between northern Lhasa block and Bangong-Nujiang suture is imaged. The results demonstrate the complexity of the crust and upper mantle structure under Tibetan plateau and agree with previous studies.

Because the process of ambient noise correlation is completely repeatable, ambient noise correlation data can be used to monitor the media velocity changes. By examine ambient noise data at different time periods between the same station pairs deployed in Malaysia, Indonesia and Thailand for over four years, significant time shifts of Rayleigh wave signals are observed after three major Sumatra earthquakes in 2004, 2005, and 2007, respectively. In addition to the time shift after those earthquakes, we observed a plausible precursor signal one month before 2004 event. The time shifts appear to be frequency dependent. The absence of corresponding time shift in Love wave suggests the signal is not from clock error. The observations are interpreted as stress changes and subsequent relaxation in upper-mid crust in the immediate vicinity of the rupture as well as the broad area near the fault zone.

*Dedicated to my parents,
Qiaorong Wei and Fen Xu,
and to my wife,
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for their love, support, and encouragement.*

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PREFACE

The complexity of the crustal and upper mantle structure in eastern Asia owes its existence to the interaction of several tectonic plates. China, geographically centered in eastern Asia, is particularly influenced by two ongoing major tectonic activities. In the southwest, the Indian plate is colliding with Eurasia plate, causing the rise of the highest Plateau in the world, the Tibetan Plateau. In the east, the Pacific plate is subducting under Eurasia plate, forming volcanic arc and marginal seas. In the Southeast Asia, several plates are interacting with each other, making Southeast Asia one of the most earthquake active areas. Because of tectonic activities, eastern Asia breed many most deadly earthquakes, including 1976 Tangshan earthquake, 2004 Sumatra earthquake, and the 2008 Wenchuan earthquake. Therefore, studying the crust and upper mantle structure will help to gain more insight on the mechanism behind tectonic activities and perhaps shed a light on earthquake prediction. In my dissertation, I am investigating the crustal and upper mantle velocity structure in eastern Asia, particularly China and its surrounding areas using different seismic tools and data.

Chapter 1 and Chapter 3 are focusing on the structure of Tibetan Plateau and its eastern and southern eastern margin. As the result of the only place of active continental-continental collision, Tibetan Plateau has fascinated every geophysicist and geologist for decades. However, fundamental problems, like how Tibetan Plateau deforms, still remain unsolved due to limited data. With the improvement of modern techniques, more and more seismic stations are deployed in this remote area. However mapping detailed structure became possible only within recent years thanks to several major international research projects (e.g. INDEPTH, Hi-CLIMB, HIMNT). Accurate velocity structure of this area is essential in justifying different deformation models and providing evidence for geological interpretation. In Chapter 2, I zoom out of Tibet to investigate shear velocity structure of China and its surrounding area. By doing so, several regional geological structures are fit into a broader geological background, therefore different geological units/blocks are related and interactions between different units/blocks are better studied. A high resolution crust and upper mantle China shear velocity model is provided in Chapter 2 that can be used as a reference model for future regional and countrywise geophysical studies. Chapter 4 switches gears from studying velocity structures to investigating temporal changes of media velocity associated with major earthquakes in Sumatra region. Seismologists have been struggling in the area of earthquake prediction. Monitoring the temporal changes

before and after earthquakes can help seismologists understand earthquake genesis and recurrence. With newly developed ambient noise correlation techniques, it is now possible to detect small velocity changes with unprecedented accuracy. In Chapter 4, I study three major earthquakes in Sumatra in 2004, 2005 and 2007. Significant time shifts are observed after these earthquakes. A plausible precursor signal is detected one month before the 2004 event. If such a signal can be detected and confirmed after each major earthquake, it would be a major breakthrough in earthquake prediction.

By far, seismological methods are still the most efficient and accurate geophysical tools in probing the earth's interior structure. Among all of these, seismic tomography is one of the most popular methods. Seismic tomography belongs to a broader category called inverse problems, which solves physical properties from direct or indirect observations. This approach is appealing in that observations (e.g. seismic wave arrival times) are relatively easy to obtain while the physical properties (e.g. velocity), which we are interested in behind those observations, are much harder to measure directly. The concept of seismic tomography was proposed long ago. But it became practical only when computers were capable of dealing with large data sets and the complicated computation. Nowadays, seismic tomography is evolving with an elevated speed thanks to the powerful workstations and super computers. However, an inherent problem of seismic tomography that seismologists have to deal with is the non-uniqueness. One of the major reasons that cause non-uniqueness is the trade-offs between different physical. The solution to this is to include different data types that are sensitive to different properties. In this dissertation, I developed two joint inversion methods trying to improve tomography resolution.

Chapter 1 describes a body wave tomography method, jointly inverting first and secondary Pg and Pn arrival data to image simultaneously crust and uppermost mantle P wave velocity as well as Moho depth. Combining Pg and Pn data solves the trade-off between Moho depth and crustal velocity. By including secondary Pg arrivals, which typically sample the lower crust, the velocity structure of the whole crust is better constrained. Therefore in return, the Pn velocity and Moho depth can be more accurately determined. In Chapter 3, I develop a nonlinear joint inversion method using surface wave dispersion and receiver function data. This data combination addresses the velocity-depth trade-off in receiver functions and the problem of insensitivity to velocity jumps in surface wave dispersion inversion. Therefore a joint inversion is able to yield

sharper Moho discontinuity than using surface wave dispersion alone with more accurate S velocity structure. In the inversion, I adopt the Neighborhood Algorithm (NA) to directly search S velocity structure and Moho depth beneath seismic stations. Unlike the traditional linear inversion that I use Chapter 1, nonlinear inversion methods such as NA have advantages. For example, most geophysical inversion problems are nonlinear problem. Therefore it is a more natural approach to use nonlinear methods. Other benefits are that nonlinear approaches are less likely to be trapped in local minima and give relatively straightforward error analysis. When parallel computations, e.g. message passing interface (MPI), are implemented, NA can take advantage of super computer to greatly reduce computational time and then becomes a powerful tool in seismic tomography.

Traditionally, seismic studies almost exclusively rely on energetic sources, e.g. natural earthquakes and explosive sources. Since manmade sources have less energy to penetrate into the earth and the cost are exceptionally high, earthquakes are the most commonly used source. This sets a great limitation in seismic studies, particularly tomography studies. Most earthquakes are concentrated at tectonic plate boundaries, and therefore the earthquake distribution can be very non-uniform. Moreover, uncertainties, e.g. earthquake location and origin time, exist in the earthquake sources. Those limitations would affect the tomography results in a negative way. During the past ten years, a new technique named ambient noise correlation emerged and quickly became one of the most popular branches in seismic studies. This method can retrieve the Green's Function of seismic waves (mostly surface waves) from correlating diffusive wavefields, i.e. ambient noise data and earthquake coda waves, recorded at two seismic stations. This method provides more uniform seismic ray coverage and doesn't rely on earthquake sources thus eliminates the source uncertainties. Therefore, ambient noise correlation has rapidly found applications in various seismic studies.

In Chapter 2, I used a large amount of surface wave data from ambient noise correlation for surface wave tomography study in China region. I requested more than 15 years of continuous data from all the available seismic stations deployed in and around China region. The ambient noise data is a major data set for China surface wave tomography. By combining traditional earthquake data, I am able to obtain unprecedented data coverage of the region for group velocity dispersion from 8 to 120 s and phase velocity dispersion from 8 to 70 s. This data set yields 3°

resolution 3D shear velocity structure of the whole region down to 150 km in depth and images remarkable features in the crust and upper mantle. I applied the same technique in Chapter 3 in Hi-CLIMB array. With about 800 km densely distributed linear array, I retrieved large amount Rayleigh group and phase velocity data from 4 to 45 s. These data are combined with receiver function data to invert high resolution S velocity and Moho depth under each station about 5 km apart.

Because the whole processes of ambient noise correlation is completely repeatable, the technique is also powerful in detecting possible temporal change of medium. In Chapter 4, I study the media velocity changes after 3 Sumatra earthquakes by searching the station pairs that have large time shifts in the periods immediately after Sumatra earthquakes. Since the station locations don't change over time, the time shifts of surface wave arrival, if they are real, are an indication of media velocity changes. In the study, I found large time shifts which are frequency dependent in the station pairs near the earthquake after the major earthquakes in Rayleigh wave. The lack of corresponding time shifts in the same station pairs in Love waves suggests the signal is not from clock error. The observations are likely caused by stress changes and subsequent relaxation in the upper-mid crust in the immediate vicinity of the rupture, as well as from the broad area near the fault zone.

From 1D velocity to 3D heterogeneity, from isotropic media to anisotropy, from travel time inversion to full waveform tomography, seismic tomography is evolving with accelerating speed. With the aid of modern computers and new mathematics, more information can be extracted from seismic records that can help us to understand better our earth.

The main part of my dissertation consists of four chapters. Chapter 1, entitled "Joint inversion for crustal and Pn velocities and Moho depth for eastern margin of Tibetan Plateau" is published on **Tectonophysics**, 491, 185-193, 2010. Chapter 4, entitled "Temporal changes of surface wave velocity associated with major Sumatra earthquakes from ambient noise correlation" is published on **Proc. Natl. Aca. Sci. USA**, 106, 14207-14212, 2009. Chapter 2 and Chapter 3 are prepared for publication and their contents were reported in 2010 AGU fall meeting.

CHAPTER 1

JOINT INVERSION FOR CRUSTAL AND P_n VELOCITIES AND MOHO DEPTH FOR EASTERN MARGIN OF TIBETAN PLATEAU

1.1 Introduction

The eastern margin of the Tibetan Plateau is marked by spectacular topographic relief, complex geology, and active deformation as one of the most active areas of continental earthquakes in the world (Fig. 1.1). The elevation changes from 200 m in Sichuan Basin and 1 km in south Yunnan to some 4 km in the plateau. The region covers several major geological blocks, including the eastern parts of the Himalaya, Lhasa, Qiangtang, Songpan–Ganzi Fold Belt, and Qaidam Basin blocks; the Himalayan Eastern Syntaxis; the southeastern margin of the TP (Yunnan, part of Myanmar) to the south; and stable and rigid Sichuan Basin and Yangtz Craton to the east. Geological, geodetic, and seismic tectonic studies (e.g., Molnar and Tapponnier, 1975; Holt et al., 1991; Wang et al., 2001) have generally indicated that the surface crustal deformation is broadly distributed throughout the plateau with eastward extrusion in eastern TP, northeastward motion in northeast margin, and southeastward to southward motion in the southeastern margin (Yunnan and Myanmar) with a clockwise rotation around the E. Syntaxis. The region is marked by several major sutures and fault systems and is seismically very active, forming seismicity belt known as the “North–South Earthquake Belt” in China. Nine $M > 7.5$ have been recorded to have occurred in the region since 1800. The great 2008 Wenchuan Earthquake (M_s 8.0) occurred in Longmenshan fault system, just west of the Sichuan Basin.

A key to understand the tectonic control of surface deformation is to gain understanding of the subsurface heterogeneous structure. Seismic imaging has been the most effective in this effort and seismic tomography is one of the widely used seismic imaging methods. Recent seismological studies in the area include receiver functions (Xu et al., 2006, 2007; Wang et al., 2008; Hu et al., 2008; Zhang et al., 2009; Lou et al., 2009), joint inversion of receiver functions and surface wave dispersions (Li et al., 2008), shear wave splitting (Lev et al., 2006; Chang et al., 2006; Sol et al., 2007; Huang et al., 2007; Ding et al., 2008; Shi et al., 2009), and wide-angle refraction surveys (Wang et al., 2007; Zhang and Wang, 2009). Several tomographic studies have also been conducted in this region in the recent years, including those using P wave travel

time tomography (Huang et al., 2002; Wang et al., 2003; Lei et al., 2009), Pn travel times (Huang et al., 2003; Liang and Song, 2006), and surface wave tomography using ambient noise (Yao et al., 2008a, b; Li et al., 2009).

In this study, we propose a new joint inversion method. A classical problem in seismic tomography is the trade-off between a discontinuity location (e.g. Moho discontinuity) and the velocity of the medium. The location of the discontinuity can affect the geometry of the ray path. Researchers commonly assume a flat Moho as a first order approximation or a priori morphology of the Moho discontinuity in tomographic inversions. However, in many regions, the structure of Moho discontinuity is not precisely known, which would result in errors in calculating travel times and ray paths. Furthermore, inaccurate Moho depth may affect the correct identification of multiple crustal arrivals and the Pn arrival. For this reason, crustal tomography and Pn tomography are typically done separately by isolating first-arriving Pg and Pn waves conservatively, by which contamination from secondary arrivals because of the existence of the Moho discontinuity can be avoided. However, lots of good data are discarded with this strategy, and, as a result, the lower crust is particularly poorly sampled.

We design an iterative scheme to invert jointly for the velocities of the whole crust and the Pn velocity as well as the Moho depth. We include first-arriving crustal P waves (Pg) sampling the upper and mid crust at relatively small distance, first- head waves Pn sampling the top of the mantle at greater distances, and secondary Pg waves sampling the mid and lower crust at distances greater than the Pg–Pn crossover distance (which is about 250 km for a 50-km thick crust) (Fig. 1.2). Note traditionally, subscript “g” in Pg indicates the P wave traveling almost entirely through the granitic layer in the upper crust (Kulhanek, 2002). Here we use Pg simply as a P wave through the crust.

The joint inversion has a few advantages. First, with different sensitivities of the phases used, the 3D crustal velocity structure, the Pn velocity, and the Moho depth can be separated and simultaneously solved in the joint inversion. Second, with accurate 2D Moho map, different phases can be better separated, thus reducing the contamination of Pg from Pn waves. Thirdly, Pn waves and secondary Pg waves can put constraint on crustal velocity, especially in the lower crust. The use of later arrivals is quite common in active source refraction seismology (e.g., Wang et al., 2007), but they are commonly ignored in earthquake crustal tomography. Because

of the complexity of the structure in the eastern margin of the TP, the region provides an excellent test case for our joint inversion method.

1.2 Method and data

1.2.1 Method

The travel time residual for Pg wave from event j to station k can be expressed in the following discrete form:

$$r_{jk} = T_{jk}^{obs} - T_{jk}^{pred} = \sum_{i=1}^n \left(-\frac{dl}{v} \right) \left(\frac{\Delta v}{v} \right)_i \quad (1)$$

where T_{jk}^{obs} is the observed travel time, T_{jk}^{pred} is the predicted travel time, and $\frac{\Delta v}{v}$ is the velocity perturbation along the ray path. The travel time residual for Pn wave can be written as

$$r_{jk} = T_{jk}^{obs} - T_{jk}^{pred} = \sum_{m=1}^{n1} \left(-\frac{dl}{v} \right) \left(\frac{\Delta v}{v} \right)_m + \sum_{r=1}^{n2} \left(-\frac{dl}{v} \right) \left(\frac{\Delta v}{v} \right)_r + \eta \Delta h_k + \Delta t_j \quad (2)$$

where n_1 is the number of grids that the seismic ray samples in the mantle while n_2 and n_3 are the number of grids that the ray samples in the crust of at the event side and the receiver side, respectively. The 1st term on the right hand side is the mantle contribution, the 2nd and 3rd term is the contribution of velocity perturbation of the crust, and the 4th term is the contribution of Moho depth variation at the receiver side. Parameter η is the vertical slowness calculated using the formula $\eta = \sqrt{\left(\frac{1}{v_c}\right)^2 - \left(\frac{1}{v_m}\right)^2}$, where v_c and v_m are the velocity in the crust at the Moho depth and the Pn velocity, respectively, at the refraction point using the updated 3D velocity model. The combination of the 3rd and 4th terms represents the station delay in the traditional formulation of Pn tomography (e.g., Hearn, 1996; Liang et al., 2004). The last term is the event delay after the removal of crustal velocity perturbation in the 2nd term, which represents the perturbation from errors in event depth and origin time, and Moho depth variation at the source side. We do not attempt to separate these contributions, which would require relocating earthquakes, and thus, this term does not contribute to resolution of the crustal structure.

Our joint inversion is a highly iterative and elaborative scheme. The gist includes rejecting bad data, sorting out different phases (Pg, Pn, and secondary crustal P), effective 3D ray tracing, formulating joint inversion for media velocities and Moho variation, iterative selection of data with the improved models, and careful treatments on the inversion process (derivation of 1D model, parameterization, regularization and smoothing, and minimization and linear-system solver). Ray tracing for Pg wave through 3D crust is done using the pseudo-bending ray tracing (PBR) technique in spherical geometry (Koketsu and Sekine, 1998). The method was recently implemented for 3D earth that includes discontinuities and an anisotropic inner core (Sun and Song, 2008). Because Pn wave is not a geometric ray, ray tracing is done approximately. We assume the great circle ray path for Pn wave. Pn ray path is divided into three segments (source to Moho, mantle part, Moho to station). We assume average crustal velocities under source and receiver in finding the Pn ray path. The model is parameterized into three sets of grids: a 3D grid for crustal velocity, a 2D grid for Pn velocity, and a 2D grid for Moho depth, respectively. Linear interpolation is used between grids.

We divide the inversion into two parts. The first part is to derive the 1D reference model, which is a simplified version of the 3D inversion described in the 2nd part. The steps of the first part are described briefly as follows. (1) We use only first-arriving Pg (less than 1°) and Pn (at $8\text{--}12^\circ$) to obtain initial 1D crustal velocity model, average Pn velocity, and average Moho depth in a 1D inversion. This is our initial 1D reference model. (2) Data with large residuals (greater than 5 s) with respect to the current 1D model is rejected. Events with depth greater than Moho depth are also rejected. (3) Associate arrivals with Pg or Pn phase. For small distances (less than 1.0°) in Yunnan–Sichuan, we calculate Pg travel time only. For large distances ($8\text{--}12^\circ$), we associate data with Pn phase only. For intermediate distances, we calculate both Pn and crustal P travel times. If Pn arrives later than Pg, we associate the phase with Pg (when the residual is less than 5 s as usual). Otherwise, we associate the datum with the phase if the travel time residual with that phase is smaller by at least 2 s than the residual with the other phase. We discard the datum if the residual difference is less than 2 s and if ray travels deeper than Moho. (4) Invert new 1D model using selected data set and repeat Steps 2 and 3 until 1D model converge. The final 1D model is used as 1D reference model in the 3D inversion.

The 2nd part of the joint inversion is to invert the 3D structure. The steps are described briefly as follows. The flow chart of 3D joint inversion algorithm is provided in appendix A. (1) Calculate predicted travel times of Pg and Pn waves with current 3D model (the initial model is the reference 1D model). Re-associate each phase accordingly following the same Step 3 above for the 1D model. Formulate inversion Eqs. (1) or (2) for each ray. (2) Due to different units used in joint inversion, coefficient matrix is preconditioned by normalizing by the column norm $\|c\|$, where c is the vector of the column and $\|c\|$ denotes the Euclidean (L2) norm with $\|c\| = \sqrt{c^T c}$. (3) We apply LSQR algorithm (Paige and Saunders, 1982a, b) to solve the problem by minimizing the following system $\min (\|Am-r\|^2 + \lambda^2 \|m\|^2 + \phi^2 \|Lm\|^2)$, where m is the parameter vector, A is the coefficient matrix, r is the vector of residuals. The first term is to minimize the data fit in the least-square sense. In the implementation, the preconditioned coefficient matrix A' is used, i.e. $A' = AC$, where C is the diagonal matrix that is the inverse of the column norm diagonal matrix, and a new parameter vector m' is inverted, which is related to m by: $m = Cm'$. The preconditioning does not change the least-square norm, i.e. $\|Am-r\| = \|A'm'-r\|$. The second term is the model regularization with a damping parameter of non-zero real number λ . With the regularization, model parameters for grids that are poorly sampled converge to the background values. This helps avoid instability. The third term is the smoothing constraint with non-zero real number ϕ . The smoothing operator L is the Laplacian operator (Lees and Crosson, 1989). (4) Model is updated with the perturbations obtained from the joint inversion. (5) Repeat Steps 1–4 until converge.

1.2.2 Data, 1D model, and model parameterization

Our data are from Chinese provincial and national earthquake bulletins as well as some handpicked arrivals in the period from 1981 to 2003 (Fig. 1.3). Raw data include 12,509 events and 250 stations distributed relatively uniformly along the margin (Fig. 1.3a). The total of raw travel time picks is 97,099. Besides first-arriving Pg and Pn waves, we use secondary Pg waves at distances greater than the Pn and Pg crossover distance. These secondary arrivals are often very energetic (see a record section in Fig. 1.3b) and are quite abundant in earthquake bulletins (Fig. 1.3c). A change of slope of first arrivals is apparent around 4° : the first arrivals at greater distances are Pn waves and we can see clear secondary arrivals at 4° to 7° , which are secondary crustal Pg waves (Fig. 1.3c). Most data are clustered around apparent travel time curves.

However, outliers are clearly visible, which apparently are wrong picks and need to be filtered out.

We carry out 1D inversion described above using the raw data to obtain our reference 1D model, sort out different arrivals, and filter out outliers. The crust is initially parameterized into vertical grids of 10 km interval with constant P wave velocity of 6 km/s from 0 to 60 km. Moho depth is set at 50 km and the initial Pn wave velocity is set at 8 km/s. Parameters (crust velocity, Pn velocity, and Moho depth) are updated after each iteration. For the purpose of tracing the crustal Pg waves, crustal velocities at the grids below Moho are extrapolated from the velocity at the last grid above Moho. 1D model converges after five iterations. Our 1D model has a Moho depth of 47.36 km, crustal velocity from 5.7 km/s to 6.3 km/s, and Pn velocity of 8.06 km/s (Fig. 1.4a).

Selected data are filtered based on the 1D model using the procedures described above are shown in Fig. 1.4b. To mitigate the problem of uneven data distribution, we further decluster these data using a procedure similar to that described in Liang et al. (2004) (in its Appendix). Essentially, the procedure finds an earthquake cluster to a station and then chooses the event-station pair whose residual is the median of all the pairs. The cluster size is set to 10 km (i.e., the distance between any two events inside the same cluster is less than 10 km). The procedure is applied to both Pg arrivals and Pn arrivals. The final dataset after the whole process includes 8167 events, 250 stations, and 31,058 arrival times. This dataset is used for inversions of the 3D model. Declustering reduces the number of similar ray paths by about 63%, thus it also improves the computation efficiency.

Our model parameterization is as follows. Crustal velocity is parameterized into 0.5° by 0.5° horizontal grids and 10 km-spacing vertical grids. Pn velocity is parameterized into 0.5° by 0.5° grids. Because of relatively sparse sampling on Moho depth (limited by station distribution), it is parameterized into 2° by 2° grids. Values between grids are interpolated linearly.

1.3 Synthetic tests

We performed a checkerboard test to evaluate the resolution of tomography images and demonstrate the effectiveness of our joint inversion method in resolving simultaneously different model parameters. The input model consists of a sinusoidal checkerboard pattern with half-wavelength of 1° and maximum $\pm 5\%$ velocity perturbation over 1D model on crustal and Pn

velocities. It also includes a Moho dipping to north from 42.5 km at 24°N to 63.5 km at 38°N to imitate the rapid change of crust thickness in Yunnan–Sichuan region (e.g., Liang and Song, 2006). Synthetic travel times are generated for the same rays as in the real inversion and a random error with standard deviation of 0.5 s is added to each synthetic travel time.

We performed two separated tests. In the first test, all the data are included (Fig. 1.5a to d). In the second test, we exclude the secondary Pg arrivals (Fig. 1.5e to h), in which case the total number of rays is reduced by 35%. In both tests, the upper crust and the Pn are well recovered. However, in the second test, the resolution in the lower crust is considerably poorer and the Moho gradient is not as well recovered. In the first test, the average Moho depths are 42.7 km and 61.8 km at 24° N and 38° N, respectively, which are within 1.7 km of the input model. In the second test without secondary P, the average Moho depth at 24° N is similar (43.0 km), but it is shallower by 3.5 km at 38°N (60.0 km) relative to the input. We conclude that our joint inversion scheme is effective in resolving the P velocities of both upper and lower crust, the Pn velocities, and the Moho. Our spatial resolution is sufficient to detect moderate scaled structures (about 1°) given our ray coverage. Lower crust structure and Moho depth variation can be better resolved with the addition of crustal secondary direct P wave arrivals.

1.4 Inversion results

In the following, we describe major features of our joint inversion results (Fig. 1.6). At shallow depths (to depth of 15 km or so), Sichuan Basin, separated by Longmenshan Fault, is well delineated by low velocity structures at shallow depth, possibly corresponding to sedimentary deposition. Another prominent feature at the shallow depths is high velocity anomalies in the Panzhihua area (Fig. 1.6a and b). The high anomaly has been identified previously (Huang et al., 2002), corresponding to mafic–ultra mafic intrusion in the area (Zhou et al., 2005). At greater depths in mid–lower crust, Sichuan basin is characterized with high velocities comparing to the plateau region to the west (Fig. 1.6c and d). In the uppermost mantle, the eastern margin of Tibet plateau shows prominent low Pn velocity surrounded by high velocity structure to the north and the east. The low and high Pn velocity structures are bounded by major boundaries and faults: Kunlun Fault to the north and Longmenshan fault, Anninghe Fault, Zemuhe Fault, and Xiaojiang Fault to the east. The Pn results have excellent agreement with previous Pn tomography by Liang

and Song (2006) with a different method and somewhat different data set, suggesting the results are robust. Moho depth is shallow in the south (~ 30 km) and in Sichuan basin (42 km or so) and dips steeply toward the north and the west (to ~ 60 km), consistent with Pn tomography (Liang and Song, 2006), receiver functions (e.g., Xu et al., 2007; Wang et al., 2008), and surface wave tomography results (Yao et al., 2008).

Fig. 1.7 compares average velocity as function of depth under Sichuan Basin (area 4) and the average velocity profiles at three sites along the TP margin (areas 1–3). We can see clearly slow upper crust, fast lower crust, and fast Pn under Sichuan Basin and slow mid–lower crust and slow Pn under the TP margin areas. We can also see rapid changes of Moho depths even in these averaged profiles.

1.5 Conclusions and discussion

We propose a new joint inversion method using first-arriving Pg, Pn waves and secondary Pg waves. The method is effective in resolving P velocities of both upper and lower crust, Pn velocities, and the Moho. In particular, lower crust structure and Moho depth variation can be better resolved with the addition of crustal secondary direct P wave arrivals. We applied the method to the eastern TP margin region, where structure is very complex. The inversion results show some major features on crustal and Pn velocities and Moho depth variation under the Sichuan Basin and the TP margin.

Of particular interest is the relatively low velocity in the mid-lower crust in the TP margin areas, compared with the average velocity profile and that in the Sichuan Basin, although no obvious P wave low velocity zones are present in the mid–lower crust in our model. On the other hand, pronounced low velocity zones in S waves in mid–lower crust have been reported for the eastern TP margin from receiver functions (Xu et al., 2007) and surface wave tomography (Yao et al., 2008). Recent surface wave tomography suggests wide spread S-wave crustal low velocity zones in the TP (Sun et al., 2008; Song et al., 2008). We compare our P model with the S model to derive the Poisson's ratio in the crust (Fig. 1.8). At shallow depth, the Poisson's ratio in the plateau region is around the average level (~ 0.25). However, in the mid and lower crust (~ 30 to 40 km depth), the plateau region shows distinct high Poisson's ratios, indicating that S velocities are more anomalous than P velocities in the mid–lower crust. This anomaly in Poisson's ratio is

consistent with receiver function study (Xu et al., 2007). These observations (low P and S velocities and high Poisson's ratio) may indicate weak or partially melting mid–lower crust in the eastern TP. Thus the mid–lower crust in this region may resemble hydraulic reservoir (Zhao and Morgan, 1987) and act as a zone for lateral extrusion and horizontal flow (Bird, 1991) or a channel for the growth of the TP margin (Clark and Royden, 2000).

We have demonstrated the effectiveness of the joint inversion methodology. However, both the method and the inversion for this region are preliminary and can be improved.

The coarse parameterization for the Moho with linear interpolation does not capture fully the rapid changes of Moho depth in this region. Using spline functions may be more effective. Earthquakes can be relocated iteratively with the 3D model inversion to improve both the structure and earthquake locations. With additional data, it's also possible to determine azimuthal anisotropy in the crust and Pn waves simultaneously, which will be important for understanding the deformation in the crust and mantle lithosphere and the coupling between them.

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1.7 Figures

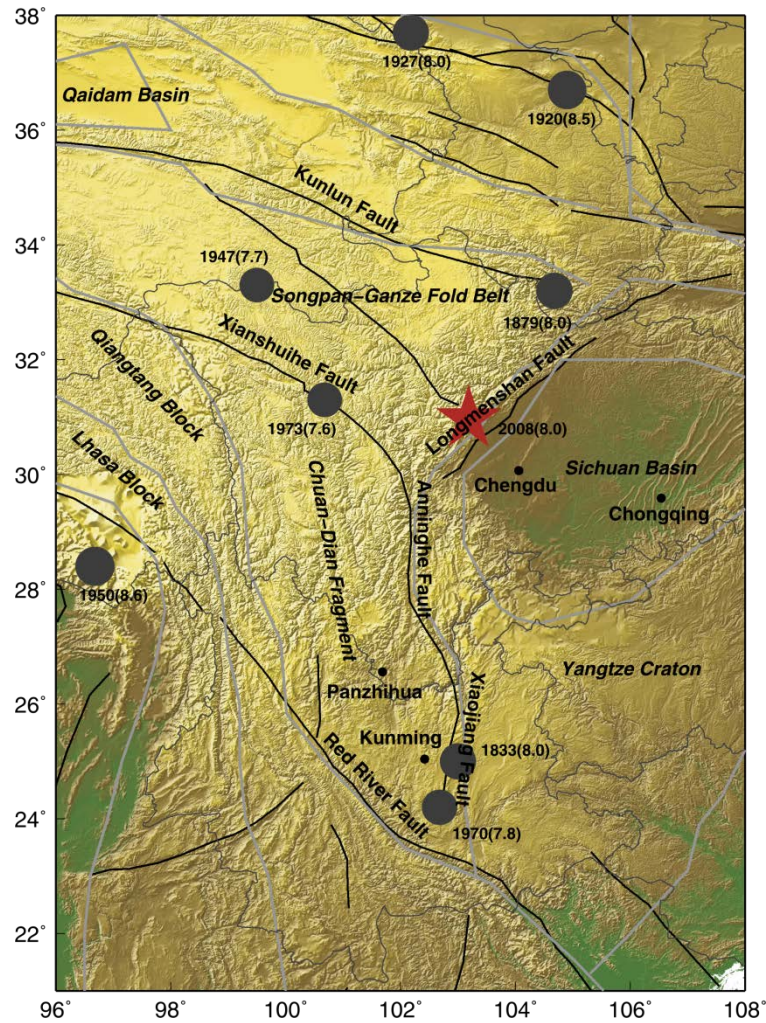


Fig. 1.1. Map of study region, including topography in the color background, major block boundaries (gray lines), major faults (dark lines), provincial boundaries (dark thin lines), and two major basins. Names of crustal blocks, faults, basins, and a few key cities are labeled. Major earthquakes with magnitude greater than 7.5 since 1800 are also plotted (the 2008 Wenchuan earthquake is indicated by the star).

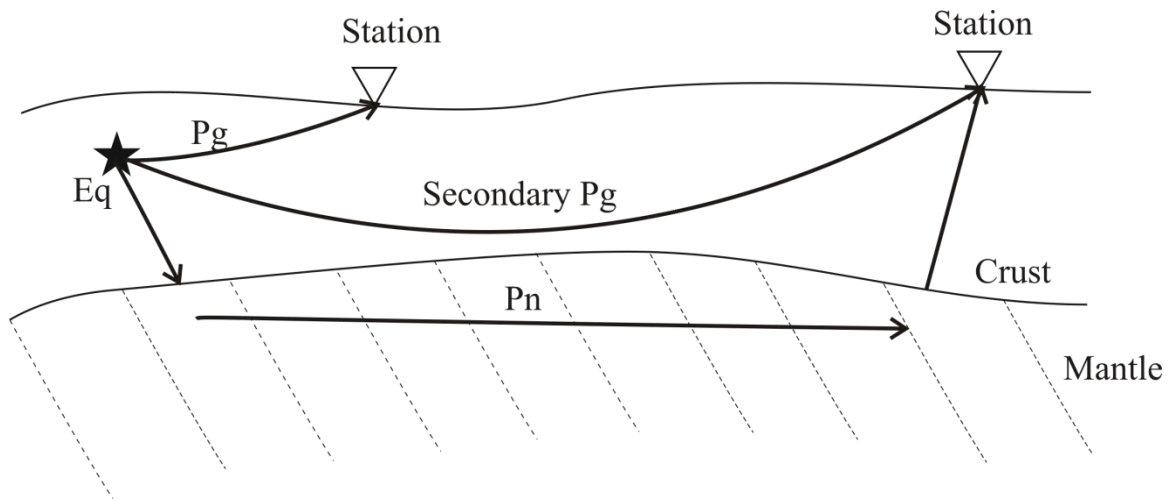


Fig. 1.2. Schematic diagram of phases used in this study. At small distances, Pg waves arrive as first arrivals sampling shallow crust. As distance increases, P waves incident upon the Moho become critically refracted at the top of the mantle and arrive at the surface as head waves. At distances greater than the crossover distance, Pn waves become first arrivals and Pg waves arrive as secondary arrivals, sampling the mid and lower crust.

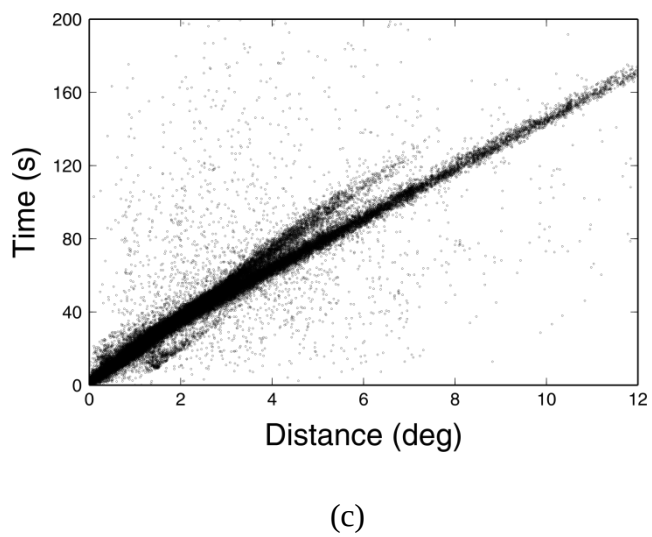
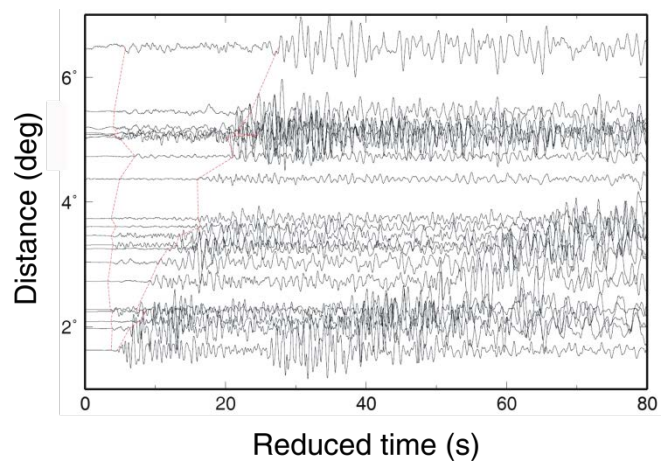
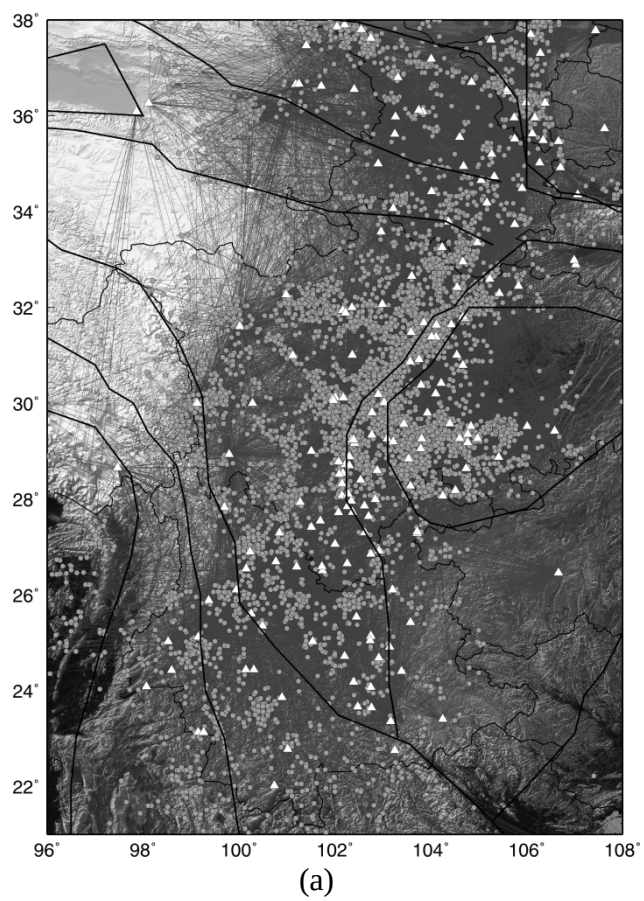


Fig. 1.3

Fig. 1.3. Raw data used in this study. (a) Ray coverage and distribution of earthquakes (light gray circles) and stations (white triangles). (b) Examples of regional seismograms at Sichuan and Yunnan stations. The travel time is reduced by 8 km/s. The lines show our travel time picks. The nearly vertical line indicates first-arriving Pn waves. The second line indicates clear and strong secondary P arrivals. Apparent velocity of secondary arrivals is about 6.0 km/s. They are Pg waves turning at mid–lower crust. (c) Raw travel time picks at stations in Eastern Tibet and margins. Data are assembled from the Annual Bulletins of Chinese Earthquakes and Sichuan, Yunnan, and Gansu provincial earthquake bulletins. Note clear secondary arrivals, after Pn, at distances from 4 to 7°.

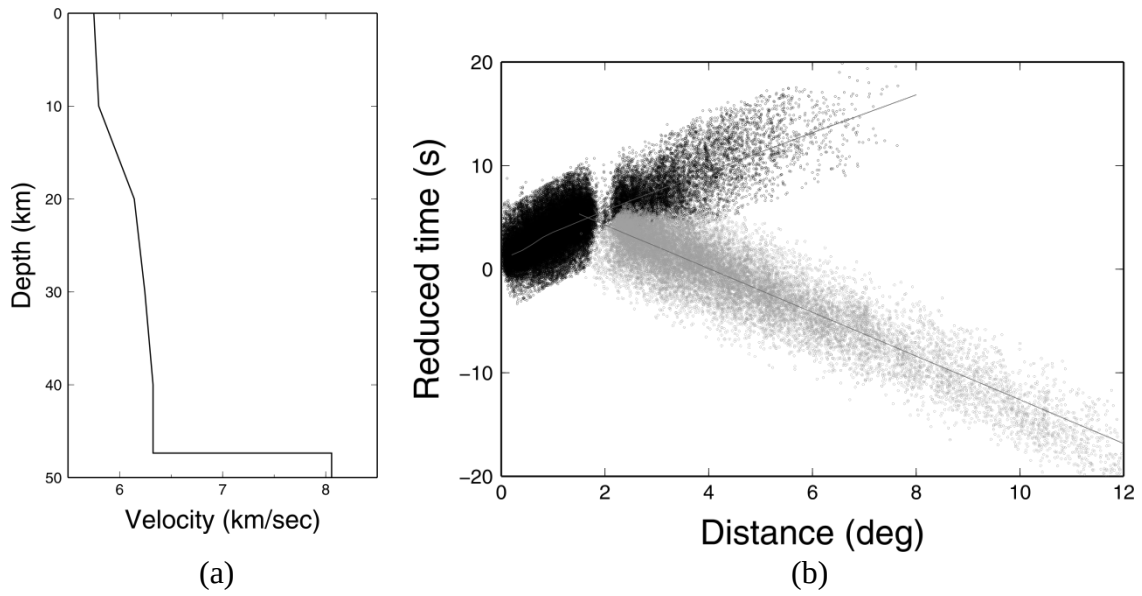


Fig. 1.4. 1D model derived in this study (a) and data filtered with the 1D model (b). In b, the travel time is reduced by 7 km/s. Light gray dots denote the picks associated with Pn phase while black dots are those associated with Pg phase. Gray lines are predicted travel time curves for Pg and Pn phases for our 1D model.

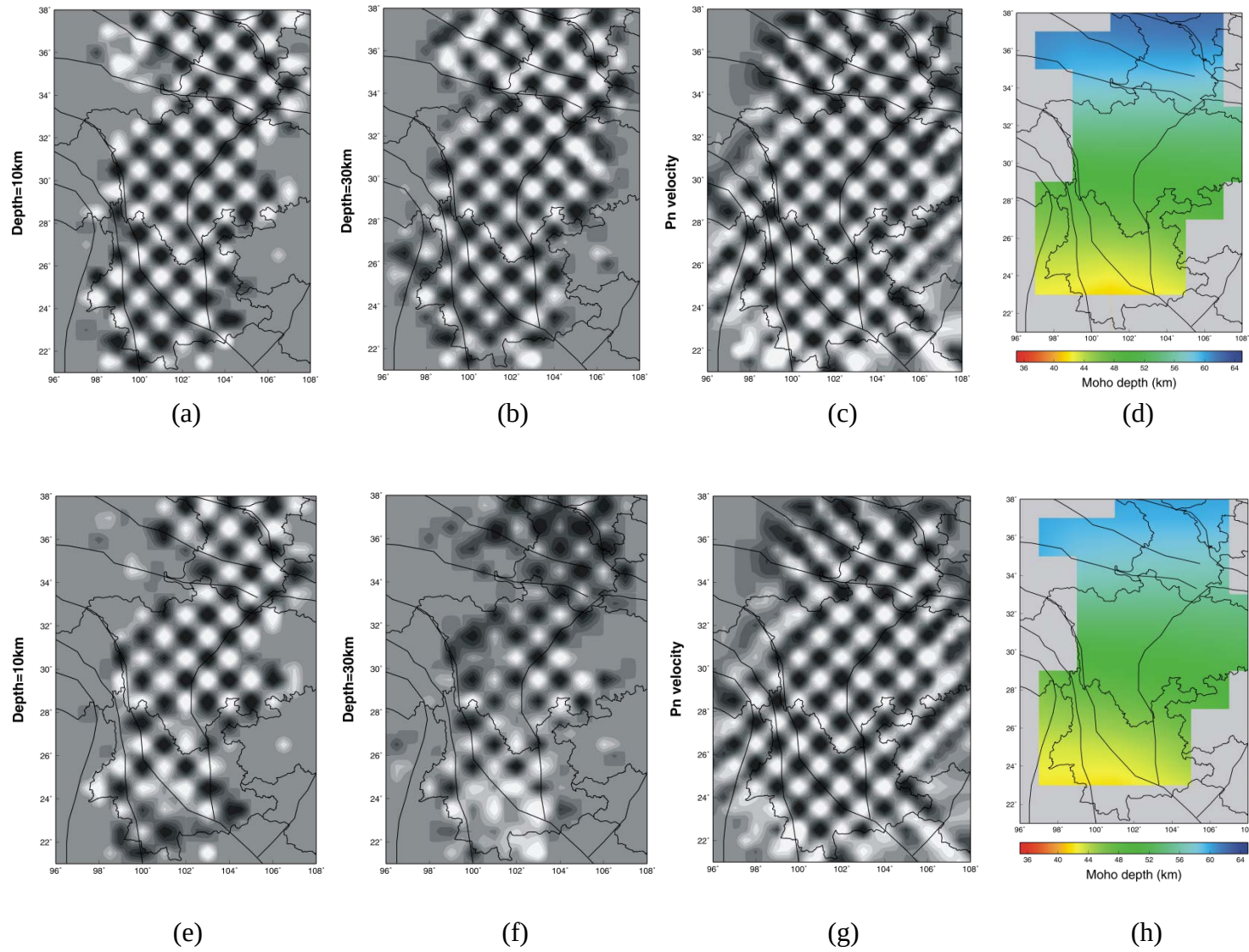


Fig. 1.5. Synthetic tests. (a–d) Synthetic test results of joint inversion using Pg, Pn, and secondary crustal P. (e–h) Same as a–d, but without including secondary crustal P waves.

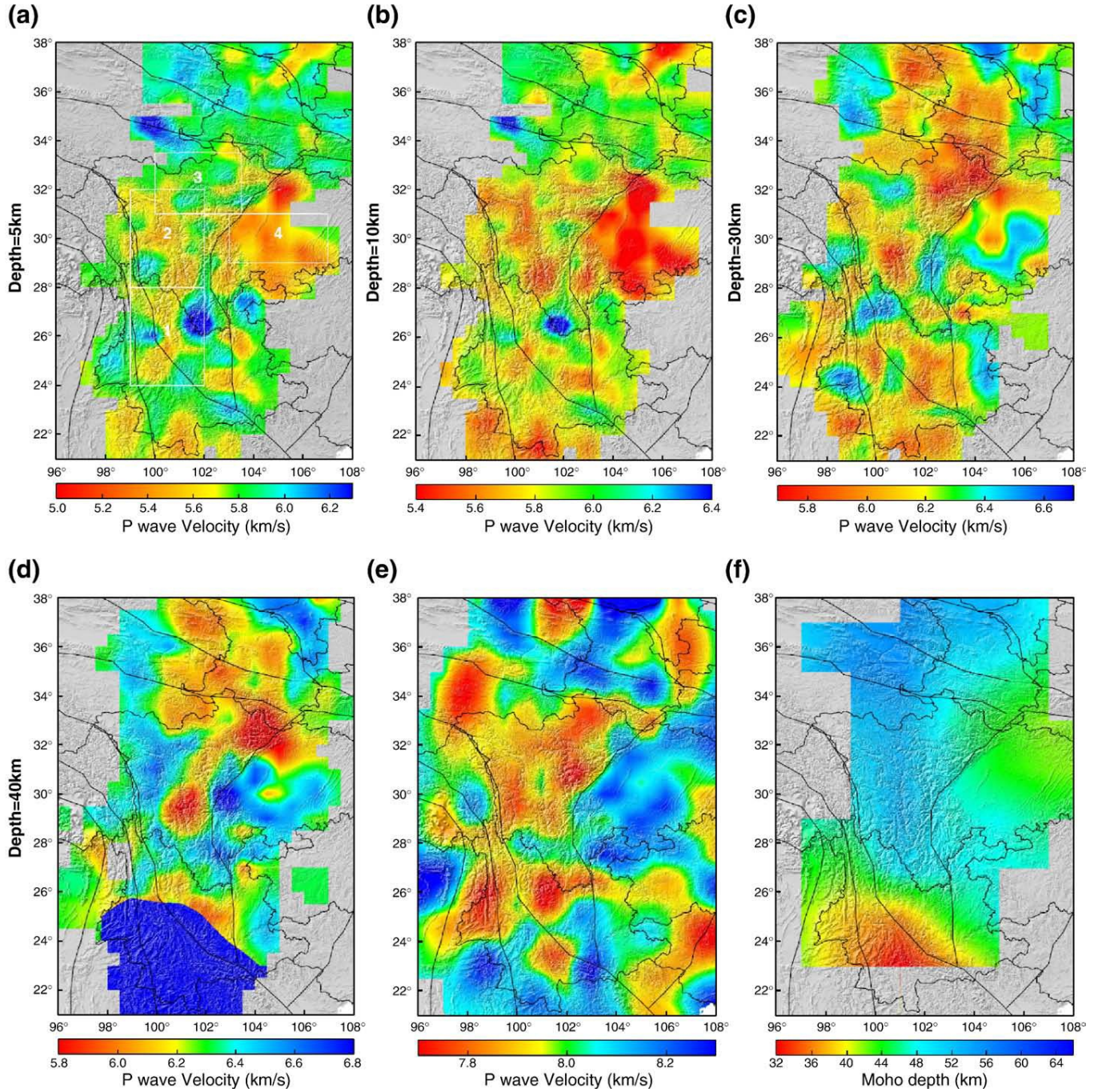


Fig. 1.6. Results of our joint inversion. Plotted are velocity at upper crust (5 km, a; 10 km, b) and mid–lower crust (30 km, c; 40 km, d), Pn velocity (e), and crust thickness (f), respectively. Areas 1, 2, 3 and 4 are referred to in Fig. 1.7.

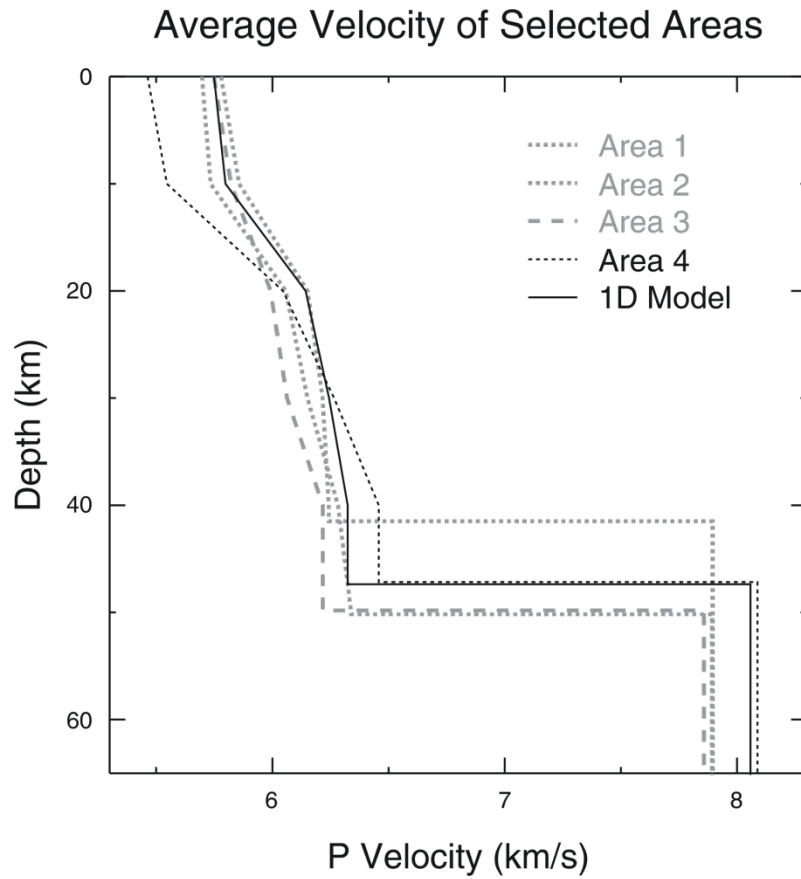


Fig. 1.7. Average velocities vs. depth for areas 1–4 marked in Fig. 1.6a. Areas 1–3 are in the plateau or southeastern margin, and area 4 is in the Sichuan Basin (see Fig. 1.6a). The 1D velocity model is also plotted for reference.

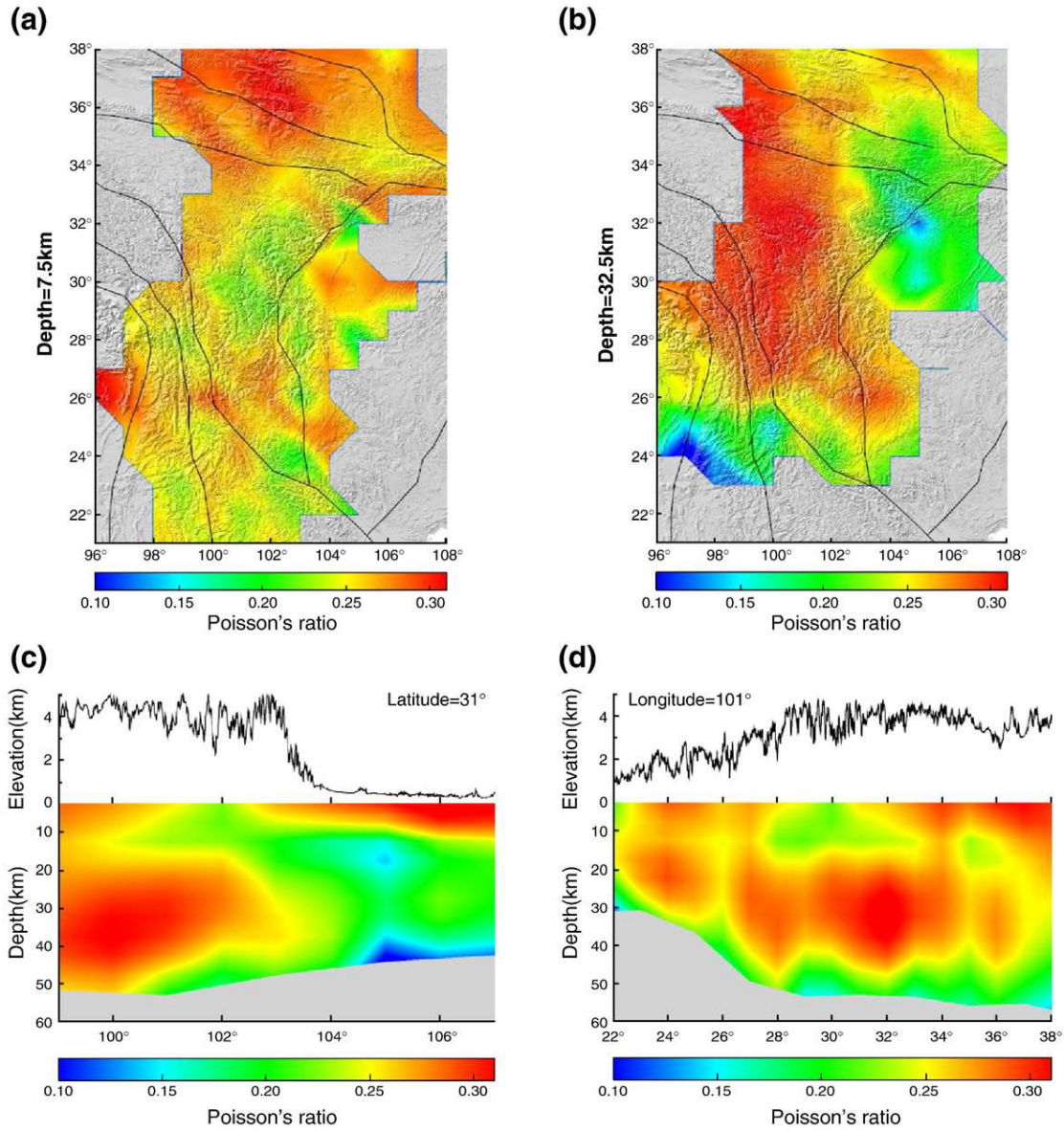


Fig. 1.8. Poisson's ratio. (a, b) Maps for depth 7.5 km and 32.5 km, respectively. (c, d) Profiles along latitude 31° and longitude 101°, respectively.

CHAPTER 2

SHEAR VELOCITY STRUCTURE OF CRUST AND UPPER MOST MANTLE IN CHINA FROM SURFACE WAVE TOMOGRAPHY USING AMBIENT NOISE AND EARTHQUAKE DATA

2.1. Introduction

Located in the southeastern portion of the Eurasian plate, China and its surrounding regions are highly geologically heterogeneous. Its diverse structures and tectonics are mainly contributed by the interaction of the Euroasian plate with the Pacific plate, the Philippine plate and the India plate. In the southeast, the India-Euroasia collision about 60 Ma leads to significant crustal shortening and the arising of the broad Tibetan Plateau with an average elevation of 5000 m, and the world's highest mountain range, the Himalaya. In the east, the Pacific plate and the Philippine plate are subducting underneath the Euroasian plate, forming the western Pacific island arcs, and marginal seas. Those aforementioned ongoing tectonics result in dramatic complexities in surface topography and crustal and mantle deformation, as well as intensive seismic and volcanic activities. Ancient platforms, deep sediment basins and high plateaus are all present in this region while separated by fault zones or mountain belts (Fig. 2.1). Due to its complexity, China and its surrounding regions constantly draw the attention of geologists and geophysicists for many years. However, there are still many interesting questions remaining unsolved, including the shortening and extension of Tibetan plateau, the rejuvenation of North China block, and the deep subducted crust in Dabie Mountain, to name a few. Therefore, investigation of the 3D crustal and mantle velocity structure will help to get a better understanding of the evolution and mechanism of these tectonic features and have important implications for earthquake hazards.

Surface waves are an important data type in studying Earth's structure. Given sparse station distribution, surface waves can provide relatively better coverage in the crust and upper mantle than body waves. In the recent decades, with the establishment of broadband digital seismic network and the deployment of more temporal seismic arrays than ever before, surface wave tomography became a more feasible tool to study the subsurface structure of China and surrounding regions. Many surface wave studies were carried out in China recently. Wu et al.

(1997) conducted Rayleigh wave tomography using group velocity from 30 to 70 s in Siberia, China, and vicinity. Ritzwoller et al. (1998) obtained Rayleigh wave group and phase velocity in Eurasia from 20 to 200 s and 20 to 125 s, respectively. Curtis et al. (1998) investigated the phase velocity of fundamental mode Rayleigh and Love wave between 20 s and 170 s in Eurasia. Villasenor et al. (2001) inverted shear wave velocity structure of central Eurasia using fundamental mode Rayleigh and Love group and phase velocities, assuming transverse isotropic model from Moho to 220 km depth. Huang et al. (2003) performed surface wave inversion of Rayleigh wave group velocity from 10 to 184 s and constructed 1° by 1° shear velocity structure down to 420 km in China and its adjacent regions. Although these studies revealed similar large-scale features in China, detailed structure information is either missing due to poor resolution or has obvious discrepancies between different studies. One important limitation of these earthquake-generated data is the uneven distribution of earthquakes and available stations, leading to non-uniform coverage. Moreover, short period surface wave, which is sensitive to shallow structures, is difficult to obtain due to the attenuation and scattering.

Recent study suggests that the empirical Green's function (EGF) of surface waves can be retrieved through cross correlation of diffusive seismic wavefield, e.g. seismic coda wave and seismic ambient noise (Weaver and Lobkis, 2001, 2004, Campillo and Paul, 2003, Shapiro and Campillo, 2004). This method provides additional data coverage with unprecedented advantages. First of all, EGFs are retrieved between station pairs, thus the ray coverage depends on station distribution, which is more uniform in most cases than earthquake distribution. Secondly, uncertainties in the earthquake source are eliminated. Furthermore, short period surface waves can be easily retrieved using this method, providing better constraint on shallow structure, especially the crustal structures. This new type of data has been quickly applied to tomography studies (Shapiro et al., 2005, Yang, et al., 2007, Bensen et al., 2008, Lin et al., 2008, Yao et al, 2008, Zheng et al, 2008, Sun et al., 2010). One drawback of this method, however, is that longer periods of surface wave signals ($> \sim 70$ sec) are usually weak and unstable. To overcome this shortage, ambient noise data are usually combined with earthquake data to provide broader period range that put stronger constraint from shallow to deep structures (Yao et al., 2006).

In this paper, we present 3D shear wave velocity results in China and adjacent regions using both ambient noise correlation and earthquake data. Part of the ambient noise data has been used in

the study by Zheng et al. (2008), and Sun et al. (2010). However, we include earthquake data and more stations that were not used in the early studies. These additional data provide denser ray coverage, and better constraints on deeper structures thus yield a more accurate shear wave velocity model.

2.2. Data and Method

The data we used can be divided into two categories: the EGFs from ambient noise correlation and the traditional earthquake data. For the EGFs calculation, we acquired 18 month continuous data in vertical component from new Chinese National Seismic Network (CNSN) in the time period from Nov., 2003 to Oct., 2004 and from Jan. 2007 to Jun. 2007. The CNSN is the Chinese backbone network that consists of 48 broadband digital permanent stations evenly covering the whole China since year 2000. We also downloaded vertical components of most continuous broadband and long period data from permanent global and regional network stations as well as temporary PASSCAL stations from IRIS DMC. The location of these stations ranges from 15° N to 55° N in latitude and 65° E to 150° E in longitude. The time span of the data is from Jan. 1991 to Jun. 2007. The total number of stations included in our study is 528 (Fig. 2.2), much larger than the preceding studies. In the earthquake data set, we gathered available event data for all the stations mentioned above from 155 earthquakes in the same region with depths shallower than 40 km and magnitudes larger than 5.5 in PDE catalog from 2000 to 2007. We also included event data recorded by CNSN stations for another 197 earthquakes with magnitude from 5.0 to 5.5 in PDE catalog in the same depth range and time period (Fig. 2.2).

All the broadband continuous data are downsampled to 1 sample per second (sps) to ensure the consistency in sampling rate. EGFs are retrieved following the procedures described in great detail by Bensen et al. (2007) and are briefly summarized here. First, the continuous data are cut into one-day-long data and bandpass filtered from 5 s to 150 s. Temporal and spectral whitening is performed after filtering process. We then compute cross correlation of seismograms for every possible station pair on a daily basis. Cross correlations of each station pair is stacked in each month. Monthly stacked cross correlation results of the same station pair are further stacked for the whole time period to obtain final EGFs. Signal-to-Noise ratio (SNR) is calculated for each stacked EGF at each period and only those EGFs with SNR larger than 10 are selected for

group and phase velocity measurement. We also require the inter-station distance should be larger than 3 wavelengths to satisfy far field approximation. Group and phase velocity dispersion curve from 8 to 70 s of each selected EGFs is measured using frequency-time analysis (FTAN) (Levshin and Ritzwoller, 2001). Group velocity dispersion curves of earthquake data are measured in the same way for periods from 8 to 120 s. We choose $-\pi/4$ as the phase ambiguity term for EGFs, following the discussion by Benson et al. (2007). However, we exclude the phase velocity measurement of earthquake data to avoid the uncertainty of initial phase. Fig. 2.3 shows group velocity measurements at different periods. The newly included earthquake data account almost 50% of total group measurements, providing significant improvement in data coverage, especially at mid to long period ranges. The incorporated data set has maximum of ~ 35000 measurements for group velocity and ~ 22500 measurements for phase velocity at period of 20 s, almost four times larger than the measurement in the study by Sun et al. (2010).

Shear wave velocity tomography is a two-step procedure. First, we invert 1° by 1° phase and group velocity at different period (8 to 70 s for phase velocity and 8 to 120 s for group velocity) using selected phase and group velocity dispersion measurements. It is a linearized 2D inversion by minimizing the following equation:

$$\min (||\mathbf{A}\mathbf{m}-\mathbf{r}||^2+\lambda^2||\mathbf{m}||^2+ \varphi^2||\mathbf{L}\mathbf{m}||^2)$$

The first term is data misfit, in which $\mathbf{m}$ is the parameter vector, $\mathbf{A}$ is the coefficient matrix. $\mathbf{r}$ is the vector of travel time residuals. The second term is the model regularization. The third term is the Laplacian smoothing constraint where $\mathbf{L}$ is the Laplacian operator (Lees and Crosson, 1989). Non-zero real number λ and φ are damping parameter and smoothing parameter, respectively. We experimented different combination of λ and φ and found there were only minor differences in velocity structures which are beyond our resolution limits, given a certain range of λ and φ . Since our resolution is about 3° by 3° (see synthetic test), we chose λ and φ so that structures larger than 3° are preserved while at the mean time small wavelength fluctuations are eliminated. Since our data contains large amount of measurements involving over 500 stations and 300 events, bad quality data still possibly exist in the data set, even data are selected according to SNR. Therefore, it is necessary to apply additional data cleaning. First, we use data from permanent stations and reliable PASSCAL stations to invert group and phase velocities for different periods. Second, we calculate theoretical travel time and travel time residual for each

measurement using derived models. In the third step, we calculate mean and standard deviation of absolute and relative residuals (residual in percentage relative to observed travel time) for each period. Last, we exclude the measurements with either absolute or relative residual outside two standard deviations from the corresponding mean value. The reason we also consider relative residual is that relative residual is less likely affected by distance. We perform above cleaning processes twice to obtain the final data set. Final group and phase velocity maps are then inverted using cleaned data.

The second step of shear wave tomography involves 1D inversion of shear wave velocity for each node at different depth from final group and phase velocities. 3D velocity maps then are constructed through linear interpolation between different nodes using GMT program. We use the computer program developed by Robert Herrmann (1991) to perform the second step of inversion. We parameterize the model into 25 layers overlying half space. Layer thickness is 5km layers on the top 50 km, 10km for the next 100 km, and 25km for the bottom 125 km. We assume homogeneous velocity model as initial model. Only shear velocity is inverted because Rayleigh wave is sensitive primarily to shear velocity structure. Poisson's ratio is fixed during inversion. P wave velocities are updated according to S wave velocity changes. Density is calculated using Nafe-Drake curve.

2.3. Synthetic test

To test the resolution of our inversion, we perform a checkerboard test. The input model consists of 3° by 3° alternating positive and negative pattern with 5% velocity perturbation. From the shear wave model, we calculate group and phase velocity at each node for the periods we used in the real inversion. We then calculate group and phase dispersion curves using exact the ray path as in real inversion. We add 5 s random noise to the travel time calculations to mimic the noise level in real data. Shear wave velocities are inverted using a two step inversion as described in the last section (Fig. 2.4). The checkerboard pattern is generally well recovered for the top 150 km in most of mainland China, even though the resolution gradually deteriorates at deeper depths. The best resolution appears in the southwestern China, namely the Tibetan Plateau and its surrounding regions due to the densest ray coverage from different PASSCAL experiments and earthquake events.

We also perform a synthetic test for the velocity model under the Tibetan Plateau. A widely spread mid-crust low-velocity zone in Tibetan Plateau and slow mantle velocity in northern Tibet are imaged from our tomography (Fig. 2.6, 2.7). Our input model consists of a low velocity layer from 20 to 40 km inside Tibetan Plateau with velocity decrease of 5%. Low mantle velocity structure is confined only in northern Tibet with a 5% velocity decrease from normal mantle. Crustal thickness within Tibet is 70 km while outside it is 40 km. We use the same forward calculation procedure as described in checkerboard test. Fig. 2.5 shows the three cross sections of input models and the recovered models. Mid crust low velocity zones and mantle low velocity structure are resolved in our cross sections, although the Moho boundary between crust and mantle is not sharp in our image, due to the lack of vertical resolution in surface wave. Nevertheless, given the size and amplitude of the anomaly, we are confident that low velocity features in the mid crust and uppermost mantle are real and can be well resolved.

2.4. Results

Our 3D shear wave velocity results shows remarkable features that correlate with surface geology and regional tectonic settings (Fig. 2.6). At shallow depth (7.5 km), low velocity structures correspond to major sediment basins, e.g. Jungar, Tarim, Qaidam, Ordos, Sichuan, Songliao, North China (locations of these basins are marked in Fig 2.1), consistent with underlying thick sedimentary layers. Low velocity also appears at Himalaya collision zone, Central to Northern Tibet, Pacific island arc and marginal seas. However, at lower crust and upper mantle depth (55 and 75 km), the velocity under Himalaya collision zone becomes faster. High velocities mainly appear in eastern China, such as South China fold belt, Yangtze craton and Korean peninsula, indicating a cool and strong upper crust.

At depth around 30 to 50 km, tomography results present a striking feature of the velocity contrast between eastern and western China. The boundary between fast and slow velocities shows a NNE-SSW trend, consistent with observed gravity lineation. The rapid transition from high velocity to low velocity across the eastern margin of Tibetan plateau coincides with the sharp changes in elevation between eastern and western China, demonstrating thickened crust in Tibetan plateau from India-Eurasia collision. As a result, at these depths, Tibetan plateau is still in the crust while eastern China is already in the mantle. The thickened crust in Tibet can be

traced to around 75 km depth, where the Tibetan plateau is delineated with low velocity structure in contrast to relative high velocities outside plateau. Another prominent feature revealed at this depth range is the low velocity structures in Tibetan Plateau. The low velocities appear in most of the central, eastern and southeastern part of Tibet with velocity decrease of up to 15%. This low velocity feature also shows up clearly in the vertical cross sections (Fig. 2.7 a-d, f,g). In spite of the low velocity in most of the Tibetan plateau, major basins in western China surrounding Tibet, including Tarim, Junnar, Qaidam, Ordos, and Sichuan basin, exhibit high velocity around 50 km. The high velocity feature extends down to about 145 km in the upper mantle, suggesting these basins are thick tectonically stable blocks.

At greater depth (95 km to 145 km), the velocity structure in Tibetan plateau shows a more complicated pattern. Low velocity appears in western and northern Tibet. The low velocity feature extends eastward to Songpan-Ganzi fold belt, and then turns south beneath eastern Tibetan Plateau then further south into Indo-China. The consistency between geological boundaries on surface and sharp velocity contrast boundaries in uppermost mantle may suggest strong deformation coupling between crust and, at least uppermost mantle. Southern Tibet, on the other hand, shows high velocity structures, indicating underplating cold and rigid Indian lithosphere. The sharp velocity transition from south to north Tibet at Bangong-Nujiang suture suggests it is the major tectonic boundary that separates the mantle structure between northern and southern Tibet. Our tomography results also revealed eastern-western velocity variation in the northern Tibet, where a high velocity feature is imaged from 92° to 98° E in the eastern Tibet. Interestingly, this high velocity feature was also found in the results of Pn tomography (Liang and Song, 2006) and Rayleigh wave tomography (Huang, et al., 2003). In the east, Yangze craton is characterized by prominent high velocity feature while South China fold belt immerses in low velocity. Ordos basin, located in western North China Craton, is also seen in high velocity. However the velocity in eastern part of North China Craton appears to be slow. The low velocity in South China Fold Belt and eastern North China Craton at this depth range suggests the sampling of asthenosphere. Therefore, the velocity contrasts in eastern China reflects different lithosphere thickness, as shown in the velocity cross sections (Fig. 2.7 f, h-k).

2.5. Discussion

2.5.1 Comparing group velocity model with earthquake data only

Since earthquake data account for about 50% of total group velocity measurements from 8 to 70 s, it is necessary to validate the quality of earthquake data. We perform group velocity inversion with earthquake data only and comparing the results to the images using the whole data set. The results from two dataset are generally compatible (Fig. 2.8). Although there are slight difference in detailed structures, both inversions show similar features: At 10 s, major basins in China are manifested by low velocities while the stable Yangtze craton and South China fold belt shows clearly high velocity. Group velocity maps at 30 s exhibit velocity contrast between eastern and western China. 50 s maps show the thickened Tibetan crust, surrounded by high velocity blocks. Velocity in eastern China is generally high. At 70 s, Rayleigh wave group velocity is mainly influenced by mantle lithosphere structure. Low velocity presents in Tibetan Plateau, eastern North China Craton and marginal seas while high velocity is shown in the major basins in western China, Yangtze craton, and South China Fold Belt.

To further test the data, we compare group velocity inversion result using ambient noise data only with that of using earthquake data. We don't include the background reference model, so that the velocity model is solely controlled by the data. Group velocity maps at 30 s (Fig 2.8 i, j) suggest when the data coverage are robust, both data yield similar results. Therefore, we conclude that earthquake data provides consistent high quality group velocity measurements as EGFs.

2.5.2 Tibetan Plateau

To better demonstrate the velocity structure in Tibet. We construct several velocity profiles. The location of these profiles is identified in Fig. 2.6 (a). Fig. 2.7 (a-c) shows three profiles cross Tibet running along longitude (c) or near N-S (a and b) direction. These three profiles revealed similar features. Clear crust thickness variations are shown in these profiles. The thickened Tibet crust is a mirror reflection of the topology, which is plotted on top of each profile. Along these profiles, the low velocity zone presents in the mid to lower crust, although the geometry varies. Fig. 2.7 (a) shows the low velocity zone connected with low velocity on the surface, while the same feature in Fig. 2.7 (b and c) appears to be confined in the mid crust. Nevertheless, all of the low velocity structure terminates near the boundary of Tibetan Plateau. The same feature can be

seen in almost all other velocity profiles that cross Tibetan Plateau (Fig. 2.7 d, f-g). The vast existence of low velocity zone suggests it may be a common feature that spreads out most part of the plateau. Our observation of low velocity zone seems support the channel flow model proposed by Clark and Royden (2000). The connection of low velocity zone with surface slow velocity may provide evidence of the extrusion of low viscosity crustal channel (Beaumont, et al, 2001). A high velocity structure is seen in the mantle in these three profiles gently moving downward from India to the southern Tibet. The high velocity ceased at Bangong-Nujiang suture reflects the underplating Indian plate. Pronounced low mantle velocity appears in Qiangtang terrain, north of Bangong-Nujiang suture (Fig. 2.7 a-c, g). The difference between north and south Bangong-Nujiang suture was widely reported by previous studies using different methods and data. Tilmann et al. (2003) revealed a subvertical high velocity zone from 100 to 400 km south of Bangong-Nujiang suture. They interpreted this mantle high velocity structure to be downwelling Indian mantle lithosphere. Brandon and Romanowicz (1986) reported their discovery of “no-lid” zone in the central Tibetan plateau from pure path phase velocity measurements of long period Rayleigh wave. Their study suggested low shear velocity in the mantle of Qiangtang Terrain immediate underneath Moho. Receiver function study using INDEPTH data shows the absence of distinct high velocity layer in the lower crust and lack of abrupt crust-mantle velocity increase in the north of Bangong-Nujiang suture (Owens and Zandt, 1997). It is the same region where inefficient Sn propagation and low Pn velocity was reported (Barazangi and Ni, 1982, McNamara, 1995). Our tomography result, in consistent with previous observations, shows a snapshot of Indian mantle lithosphere at current stage. It is however debatable how the Indian lithosphere evolved in ancient time. Some researches support the subduction model, in which the underplating Indian plate stopped northward advancing at Bangong-Nujiang suture then subducted subvertically (Tilmann et al., 2003, Kosarev et al., 1999, Li et al., 2008) and eventually, due to mantle convection, detached and sank into deeper mantle. Other studies suggest delamination model, where underplating Indian lithosphere may advance further into Qiangtang terrain then detached into mantle due to convective instability (Chen and Tseng, 2007, Huang and Zhao, 2006). However our result is limited by the resolution to 150 km. Velocity image at greater depth is needed to further confirm or rule out the different hypotheses.

The profile in the eastern margin of Tibetan Plateau crosses the triangle Songpan-Ganzi terrain (Fig. 2.7d). Like in Qiangtang terrain, Songpan-Ganzi also shows low velocity in the mantle.

Receiver function study in this region suggested high Poisson's ratio of 0.33 in the crust due to unusually low S velocity in the lower crust (Owens and Zandt 1997). Songpan-Ganzi terrain is the region also found with insufficient Sn propagation and low Pn velocity (Ni and Barazangi, 1983, McNamara, 1995). Consistent low velocity in the crust and mantle in Qiangtang and Songpan-Ganzi terrain indicates partial melting probably related to high temperature in the underlying mantle.

Profiles run across major basins around Tibetan Plateau in the western China show similar features: Low velocity on the top reflects thick sedimentary deposition. Continuous high velocity extending down to at least 140 km suggests these basins have cold thick root (Fig. 2.7 b, f-h). Therefore, these basins are probably isolated rigid blocks that play an important role in confining the deformation of Tibetan plateau in a triangle shape. Some studies suggested subduction of Asian lithosphere under northern Tibet (Willett and Beaumont, 1994; Kind et al., 2002, Kosarev et al., 1999). However, we don't observe southward dipping structure in northern Tibet. Therefore our model doesn't seem to support this model.

Fig. 2.7 (e) shows the east-west velocity profile across eastern Himalaya syntaxis. A mantle high velocity is distinguished in eastern Himalaya. P wave tomography by Li, et al. (2006) revealed a steeply subducting Burma microplate in Indo-Burma range. A similar feature is imaged by Huang and Zhao (2006). Ni et al. (1989) studied the geometry of Burma Wadati-Benioff zone. They found the Burma microplate is descending eastward at an angle of 50° beneath Eurasian plate in the north near eastern Himalaya syntaxis and reaches a depth of about 140 km near 95° E. Our S velocity image shows horizontal velocity contrast near Indo-Burma region at depth from 120 km to 150 km, where the eastern part shows a slower mantle velocity than the western part. The thickened high velocity structure in western Indo-Burma region may be an indication of subducting Burma microplate. However, velocity images at greater depth would be critical to determine the shape of the subducting plate.

The mantle high velocity feature in eastern Tibet at around 94° in Fig. 2.7 (f) is intriguing. A similar feature is also found in surface wave tomography (Huang et al. 2003) and Pn study (Liang and Song, 2006). The reason for the high velocity structure in the eastern TP is not well understood. A geochronology study by Chung et al. (1998) suggests that the lower lithosphere under eastern TP, where the high velocity mantle structure is imaged, was removed 40 Myr ago

inducing rapid uplift of TP and magma activities. While in western TP, the removal of lower lithosphere started only 20 Myr ago. Therefore the high velocity in the eastern Tibet may correspond to the region with older magmatic age. However, we cannot rule out the possibility that the high velocity is the remnant mantle lithosphere. A better understanding of the evolution of Tibetan Plateau is needed to explain the origin of this high velocity feature.

2.5.3 North China Craton

North China Craton (NCC) is another complicated area in our study region with some large intraplate earthquakes. The NCC consists of western (Ordos Basin) and the eastern block along with the central block in between. The eastern NCC can be further divided into a southern part and northern part (Bohai Bay) (Huang et al., 2009). Geophysical observations suggest the western block is a stable part of the craton with a thick mantle root, low heat flow, and has experienced little internal deformation since the Precambrian (Kusky et al., 2007). In contrast, the eastern NCC is unusual as a craton for the present of numerous earthquakes, high heat flow and thin lithosphere (Kusky et al., 2007).

Our results show significant difference in crust and mantle velocity between eastern and western NCC. To better demonstrate the difference, we plot four profiles along latitudes and longitudes in the north, south, west, and east of NCC (Fig. 2.7 h-k). Profile (h) passes through Ordos Basin, central NCC, Bohai Bay and North Yellow Sea. At shallow depth, thickened sedimentary deposition with slow velocity is seen in the basins (Ordos, Bohai Bay). Crust gradually thins from about 50 km under Ordos basin to 35 km under Bohai Bay. In the mantle, thick root appears in western NCC. There is a sudden change in mantle lithosphere when entering the central NCC. Low velocity shows up under the thin mantle lithosphere in central NCC and Bohai Bay. Velocity under the Tanlu Fault seems to be complicated with slow velocity in uppermost mantle. Velocity difference between the western and eastern Tanlu Fault suggests it may be a trans-lithospheric fault (Huang, et al., 2003) that cuts the mantle lithosphere. However, resolution deteriorates at greater mantle depth around Tanlu Fault. Therefore, a higher resolution image is needed to discuss detailed structure under Tanlu Fault. A similar crust and mantle feature is seen in profile (i) in southern NCC, where changes in crust and mantle lithosphere thickness are well imaged. In longitude profiles, thickened roots are observed beneath two stable blocks (Yangtze

and Ordos) in western profiles (Fig. 2.7 j), while mantle lithosphere under South China Fold Belt and eastern NCC is significant thinner with thickness about 75 to 100 km (Fig. 2.7 k).

Geochemical studies of peridotite xenoliths in eastern NCC indicate the mantle lithosphere was thick (~200 km) and cold in Paleozoic time (Zheng, et al., 2007) while peridotites from basalt eruptions in Cenozoic time suggests a thin (less than 90 km) and hot lithosphere beneath the same area (Zheng et al. 2007.). These geochemistry observations suggest more than 100 km cratonic lithosphere may be removed or strongly modified in Mesozoic time (Zheng, et al. 2007). The geochemistry hypothesis is supported by our tomography result. The fast mantle lithosphere root in western NCC is at least 150 km thick while in the east, the same high velocity feature can be only traced to 70 to 100 km, which is consistent with other observations.

2.5.4 Derived Models

i) Derived Moho depth

The S velocity model we have derived can be used to explore the model characters and tectonic implications. Because of the apparent correlation of the S velocities and crustal thickness, it would be useful to derive a crustal thickness model based on the Vs model. If we obtain the Vs that corresponds to the depth of the Moho as defined by the reference model CRUST 2.0, we find that the average of the S velocities is nearly constant (Fig 2.9). We thus fit the S velocity data with a straight line. The linear trend is used as a calibration to find the true Moho (The linear trend has nearly the same Vs value with only a small positive slope). Thus to find the true Moho for any given point, we follow the steps: First, we find the reference Moho depth from CRUST 2.0. Next, we find the reference Vs from the linear trend. Lastly, we find the depth that corresponds to the reference Vs (and close to the CRUST 2.0 depth). The procedure ensures that the newly derived Moho map (Fig. 2.10a) is similar to the CRUST 2.0 (Fig. 2.10b) overall, but the new map shows much more detailed variations that correspond to variation in the S velocities. Our derived Moho depth map reflects the fundamental tectonic setting in China: The Moho depth map shows a clear thin crust in eastern China and thickened crust in western China. The Moho contrast between eastern and western China is similar to the velocity map at 30 km. The thickest crust is confined to the Tibetan Plateau, in particular the Lhasa terrain and Qiangtang terrain in

central Tibet as a result of India-Eurasia collision. In the NCC, the crust thins out toward east, similar to the changes of mantle lithosphere as shown in Fig. 2.7.

ii) Average S velocity

Since the Moho depth is significantly different between eastern and western China, comparing average velocities in upper, mid and lower crust, as well as mantle has greater implications for tectonic significance than direct comparing velocities at certain depth. We construct the average upper, mid, and lower crustal velocities and velocities in the upper 80 km mantle (Fig. 2.11, a-d). The upper, mid, and lower crust are defined as the top 1/3, mid 1/3 and bottom 1/3 of crust thickness, derived from previous step. The average upper crust velocity is very similar to velocities at 7.5 km (Fig. 2.6a), where low velocities are shown in the major basins, marginal seas and Himalayan front, while high velocity in South China fold belt. A low velocity layer in the whole Tibetan plateau is revealed in the average mid crust velocities. The eastern NCC and South China fold belt also have low velocities. In contrast, the Sichuan basin, Ordos basin and South yellow sea shows high velocity structure. The average lower crust velocity is rather flat with less variation than that in the upper and mid crust. However, there is a strong fast velocity anomaly in north western China, which is difficult to explain. This velocity anomaly also appears at 30 and 50 s group velocity maps with or without EGFs data (Fig. 2.8c-f). It is also imaged in the S model by Sun et al. (2010) using ambient noise data only. It might be an artifact due to relatively poor coverage in this area or some problematic data. However, since it is imaged from different data sets, this anomaly may also indicate real geological feature. Further investigation is needed to explore this issue in the future. Prominent features are shown in the average uppermost 80 km mantle, including high velocity in Tarim, Ordos, and Sichuan basins. Tibetan Plateau shows complicated structure with fast velocities in the Indian plate and Lhasa block, while slow velocity appears in Songpan-Ganzi Fold Belt, western and northern Tibet. Although there is some detailed difference, our mantle structure shows significant agreement with the Pn model derived from Liang and Song (2006) (Fig. 2.11e). Low velocities are also seen in eastern NCC and the South China Fold Belt as discussed in the previous velocity maps and cross sections.

2.5.5 Comparison with previous studies

Several body wave and surface wave tomography studies have been carried out in China and its surrounding area (Huang et al., 2003, Huang and Zhao, 2006, Liang et al., 2004). More regional studies have been done in the geological hot spots like North China Craton and Tibetan Plateau (Huang, et al., 2009, Zhao, et al., 2009, An., et al., 2009, Li., et al., 2006). Despite different data sets, coverage, and methods, our results show considerable similarity in major features as the results by other study groups. Fast mantle velocities in Sichuan, Qaidam, Tarim and Junggar are reported by different studies (Liang et al., 2004, Huang and Zhao, 2006, Huang et al., 2003). The model of Huang et al. (2003) shows a slow velocity anomaly in the mid crust of Tibet in a vast area between 30°-35° N and 82°-100° E, similar to the region where we find a low velocity zone. The slow and fast velocity contrast in the upper mantle between north and south of Banggong-Nujiang suture is also revealed by Huang et al. (2003). This feature is found by many research groups using different data and methods (Tilmann, et al., 2003, Owens and Zandt, 1997, Liang and Song, 2006). The low mantle S velocity region in the northern Tibet corresponds well with low Pn velocity region (Liang and Song, 2006), where McNamara et al. (1995) found inefficient Sn propagation. The cross sections and 100s Rayleigh wave maps of Huang et al. (2003) revealed relative high velocity structures between 94° and 98°. This velocity anomaly coincides well with the fast velocity structure in our S velocity model. This structure difference between eastern and western Tibet is also reported by Liang and Song (2006).

The systematic change of velocity structure from west to east North China Craton in our model is consistent with many other surface wave tomography studies (Huang et al., 2003, Huang et al., 2009, An. et al., 2009), Pn wave tomography (Liang et al., 2004) and receiver function study (Chen et al., 2009), which all suggest thinning of lithosphere and decrease in mantle velocity from west to east. The model by Huang et al. (2003) indicates lithosphere thickness decreases from 120 km in Ordos basin to less than 80 km in eastern North China Craton. This result agrees well with our model. The velocity structure near Tanlu Fault seems intriguing. Although some studies suggest the Tanlu Fault is a major fault in the eastern China that cuts across the lithosphere (An et al., 2009), detailed structure near the Tanlu Fault is still controversial. Our images of velocity structure near Tanlu Fault are not resolved by other studies. However, those studies have greater depth ranges (An et al., 2009, Tian et al., 2009, Xu and Zhao, 2009). Therefore, direct comparison with these models is difficult.

Our velocity model in southeast China shares similar features with that from Huang et al. (2003). The mantle velocity contrast between eastern part of Yangze Craton (Sichuan basin) and western part of Yanze Craton (also called Lower Yangze Craton) is clear in the map view as well as cross sections. They also reported thin lithosphere under South China fold belt about 80 km, consistent with our results. The thin lithosphere structure across different geological units in eastern China are similar (Fig. 2.7k), therefore may imply these structures are involved in the same recent tectonic processes, possibly the subduction of the Pacific Plate (Huang et al., 2003).

2.6. Conclusion

(1) We acquired a large amount of continuous and event data. EGFs retrieved from ambient noise correlation are combined with earthquake data to provide dense coverage throughout the whole area. Two step inversions were conducted to resolve S wave velocity down to depth of 150 km.

(2) Major basins are delineated in slow velocity at shallow depth, indicating thick sediment deposition. Major basins in the West (Tarim, Tulufan, Junggar, west Ordos, Sichua) as well as the Indian Shield show fast mantle lithosphere velocities, indicating strong, cold, and stable continental blocks. These strong blocks thus seem to play an important role in confining the deformation of the Tibetan Plateau (TP) to be a triangular shape.

(3) The mantle lithosphere of the northern Tibet has generally lower velocities than in the southern Tibet, correlating with the intrusion of the India lithosphere. However, there is significant E-W variation as well. The low velocity structure in the upper mantle under TP correlates with surface geological boundaries. The consistency of surface geology structure with upper mantle velocity structure suggests deformation may be coupled between crust and upper mantle.

(4) A widespread, prominent low-velocity zone is observed in the mid-crust in much of the TP, but not in the margin areas, suggesting a weak (and perhaps mobile) mid-crust.

(5) The velocity contrast between the eastern and western North China Craton is striking. The western NCC appears to be stable block with thick continental lithosphere root while eastern

NCC shows slow mantle velocity and thin lithosphere. Our results confirm the theory of reactivation of eastern NCC.

(6) P-tomography and S-tomography allows us to derive a new Moho map, which refines greatly the Moho variation relative to existing models of the whole of China. We also propose new averaged S velocities for upper crust, mid crust, lower crust, and uppermost mantle. These results provide a fundamental data set for continental dynamics and evolution.

2.7 References

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2.8 Figures

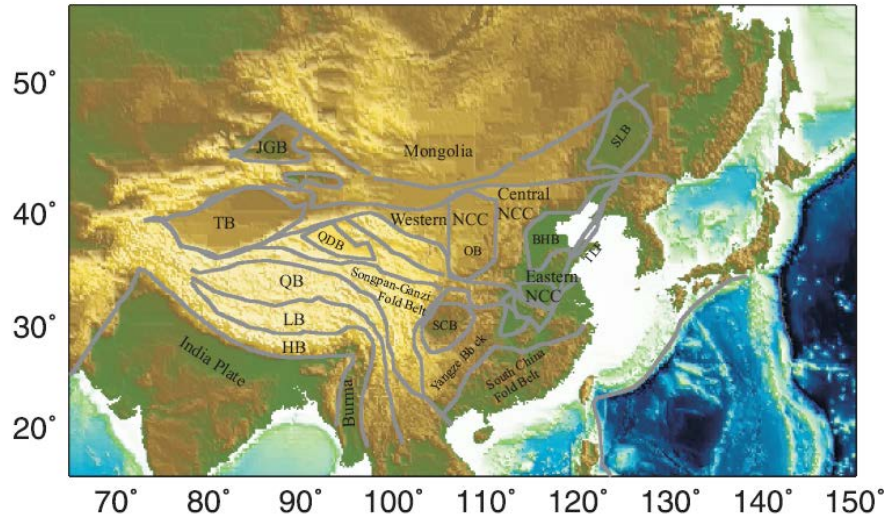


Fig. 2.1. Basic tectonic map of China. Grey thick lines indicate major geological unit boundaries. JGB (Junggar Basin), TB (Tarim Basin), QDB (Qaidam Basin), QB (Qiangtang Block), LB (Lhasa Block), HB (Himalaya Block), SCB (Sichuan Basin), OB (Ordos Basin), NCC (North China Craton), SLB (Songliao Basin), BHB (Bohai Bay), TLF (Tanlu Fault).

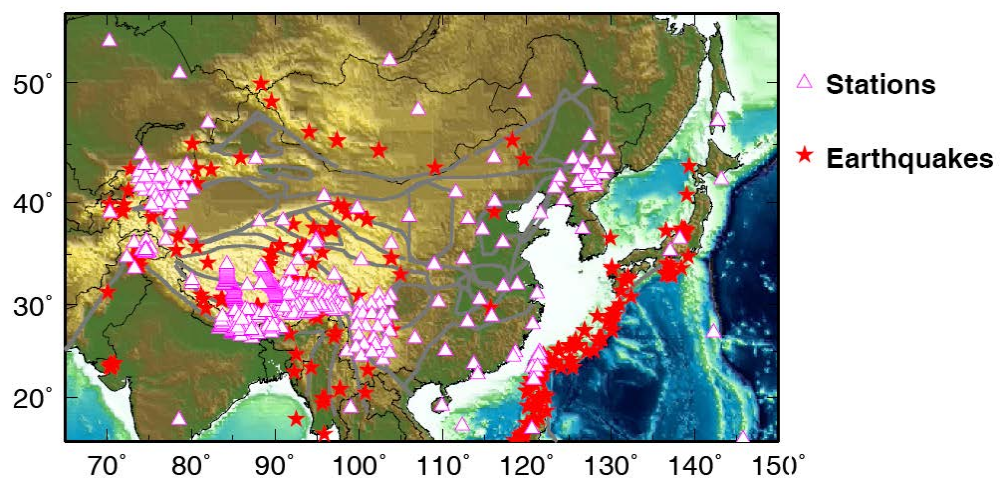


Fig. 2.2. Station and earthquake distributions. White filled triangles denote the stations that have at least one day of continuous data. Red stars indicate earthquakes from which we acquired data in our study.

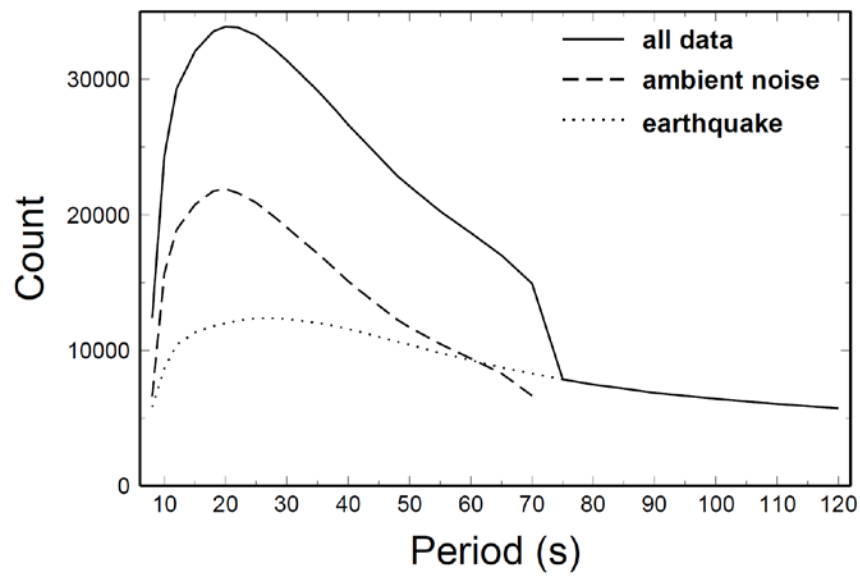


Fig. 2.3. Measurements of group velocities for different data sets. Thick dashed lines are the group velocity measurements from ambient noise correlation. Thin dashed lines are the measurements from earthquake data. Solid lines indicates the group velocity measurements for combined data set. Note only earthquake data are used for group dispersion measurements for periods longer than 70 s.

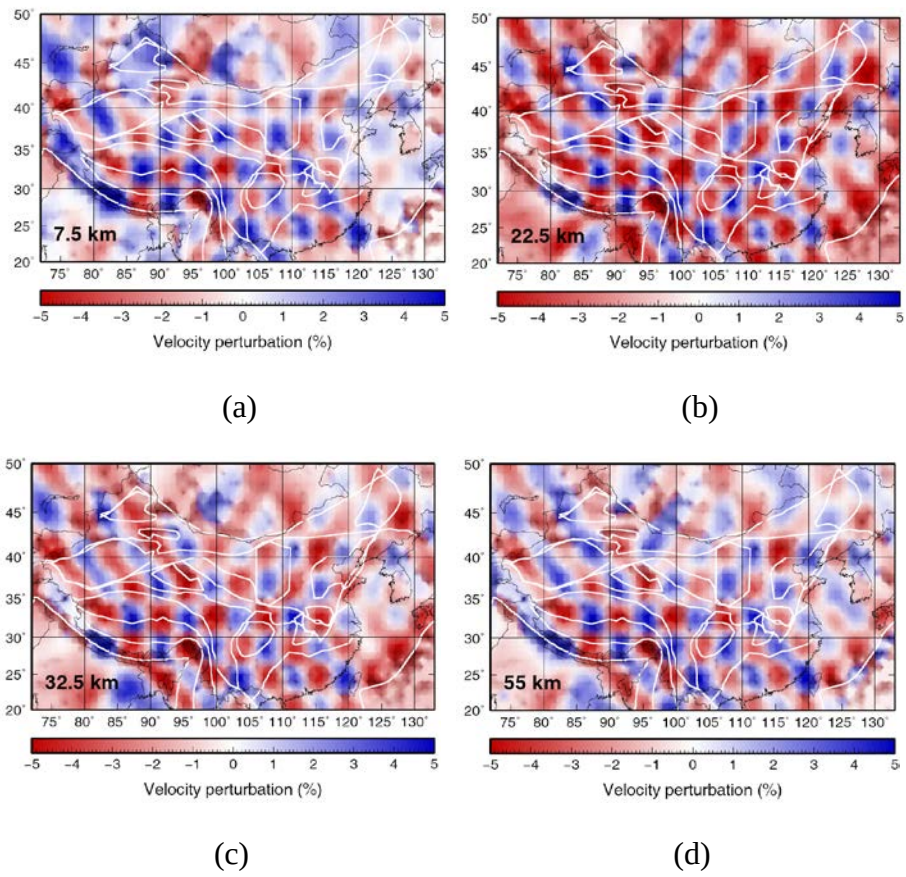


Fig. 2.4

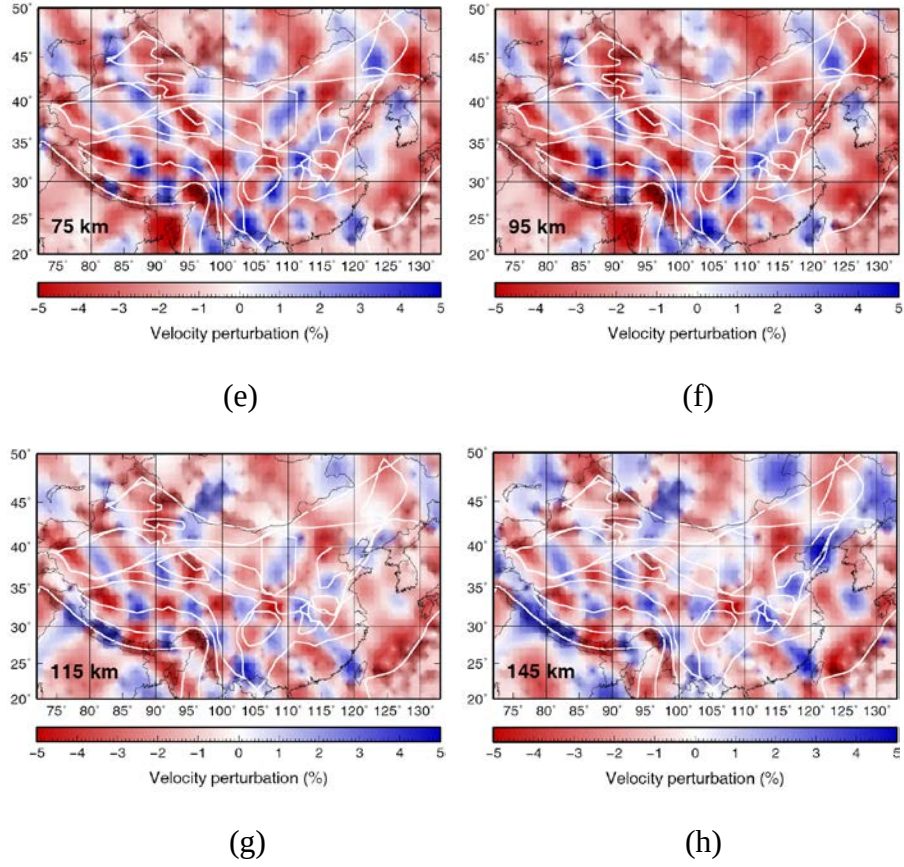


Fig. 2.4. Results of checkerboard test at corresponding depth. The input is alternating pattern of 3 by 3 degree with S velocity perturbation of 5%. 5 s random noises are added in the forward travel time calculation.

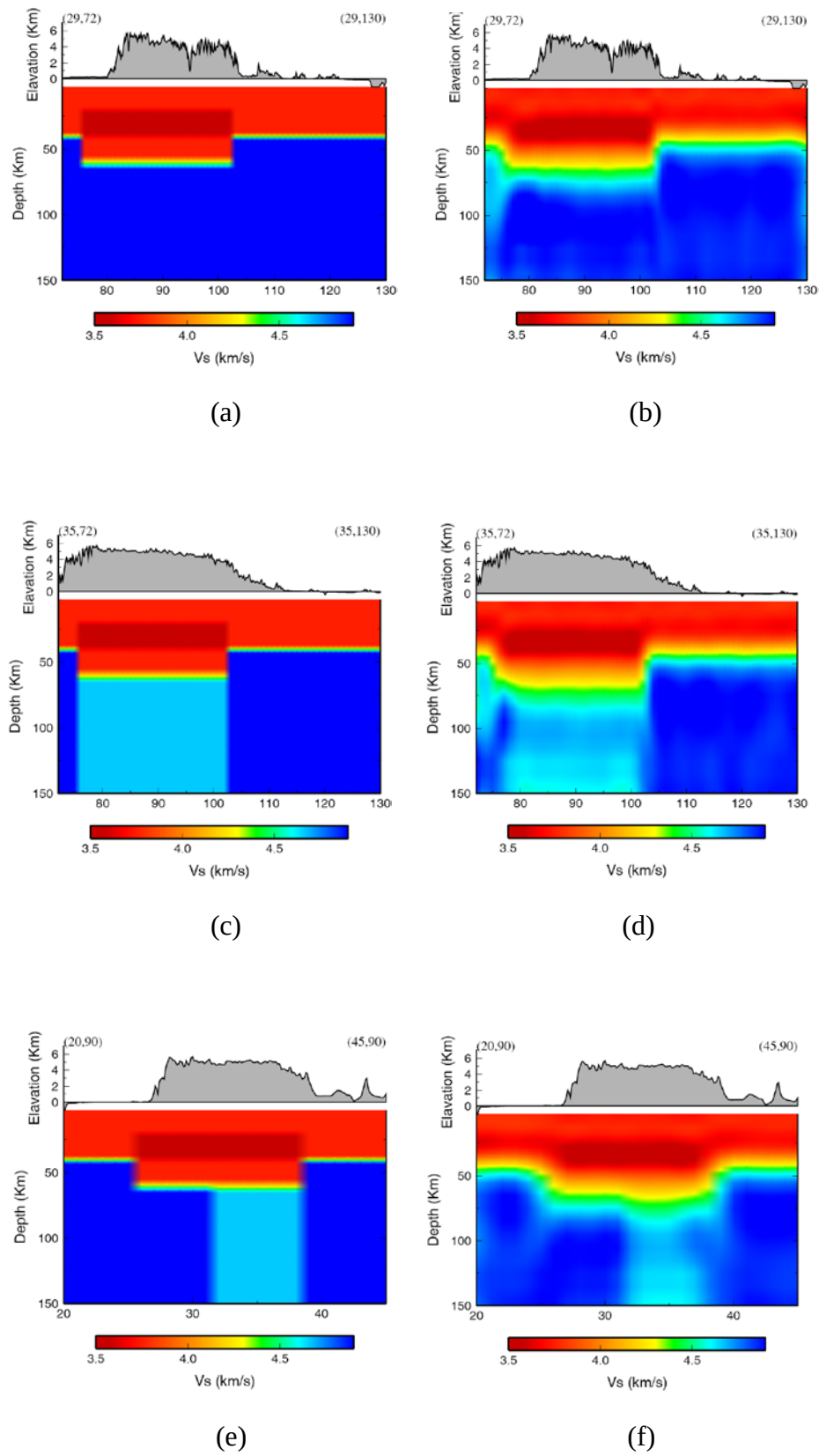
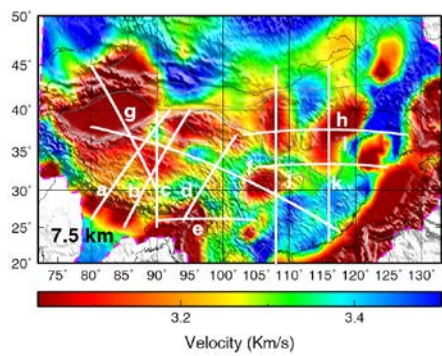
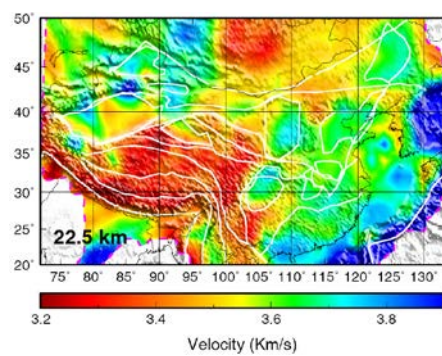


Fig. 2.5

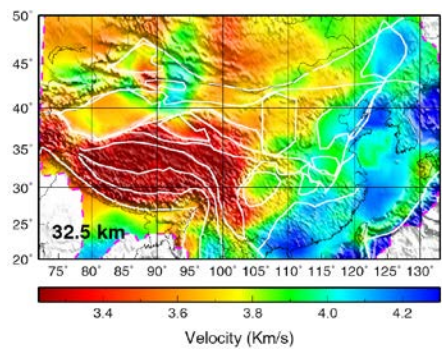
Fig. 2.5. Synthetic test of mid crust low velocity zone and slow mantle velocity in Tibetan Plateau. Figure shows three typical cross sections along 29° latitude in southern Tibet (top), 35° latitude in northern Tibet (middle), and 90° longitude across southern and northern Tibet (bottom). Cross sections of input model are shown in the left panels (a, c, e). Corresponding recovered cross sections are shown in the right (b, d, f). Note although the Moho boundary is not sharp in the recovered cross sections, the low velocity features are well recovered.



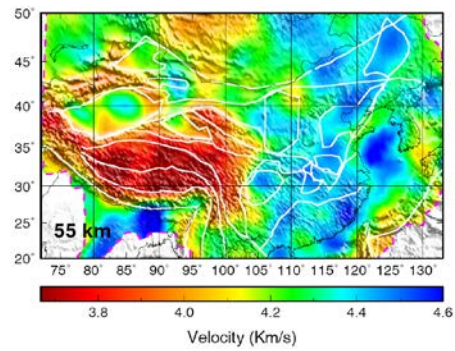
(a)



(b)



(c)



(d)

Fig. 2.6

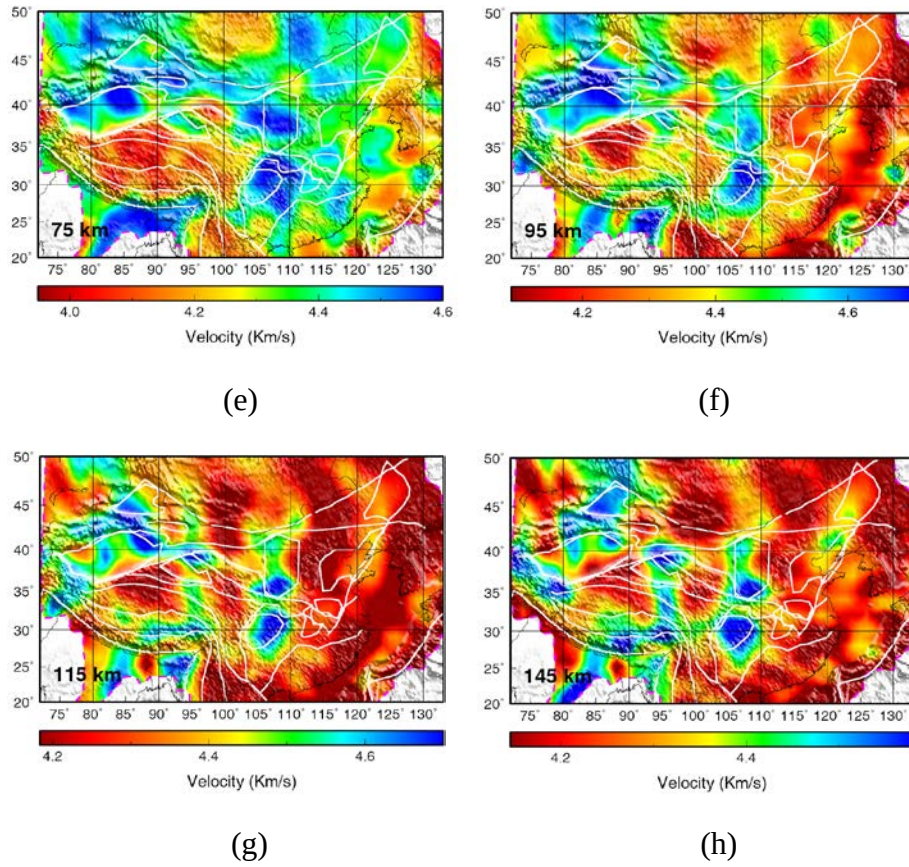


Fig. 2.6. S velocity maps at different depth. Images show remarkable features in our studying area, including sediment basin in shallow depth, mid crust low velocity zone in Tibetan Plateau, east-west velocity contrast at 32.5 km, underplating Indian lithosphere under Tibetan Plateau, Rigid continental root under Ordos, Tarim, Sichuan Basin. White great circle lines in (a) indicates the cross sections shown in Fig. 2.7.

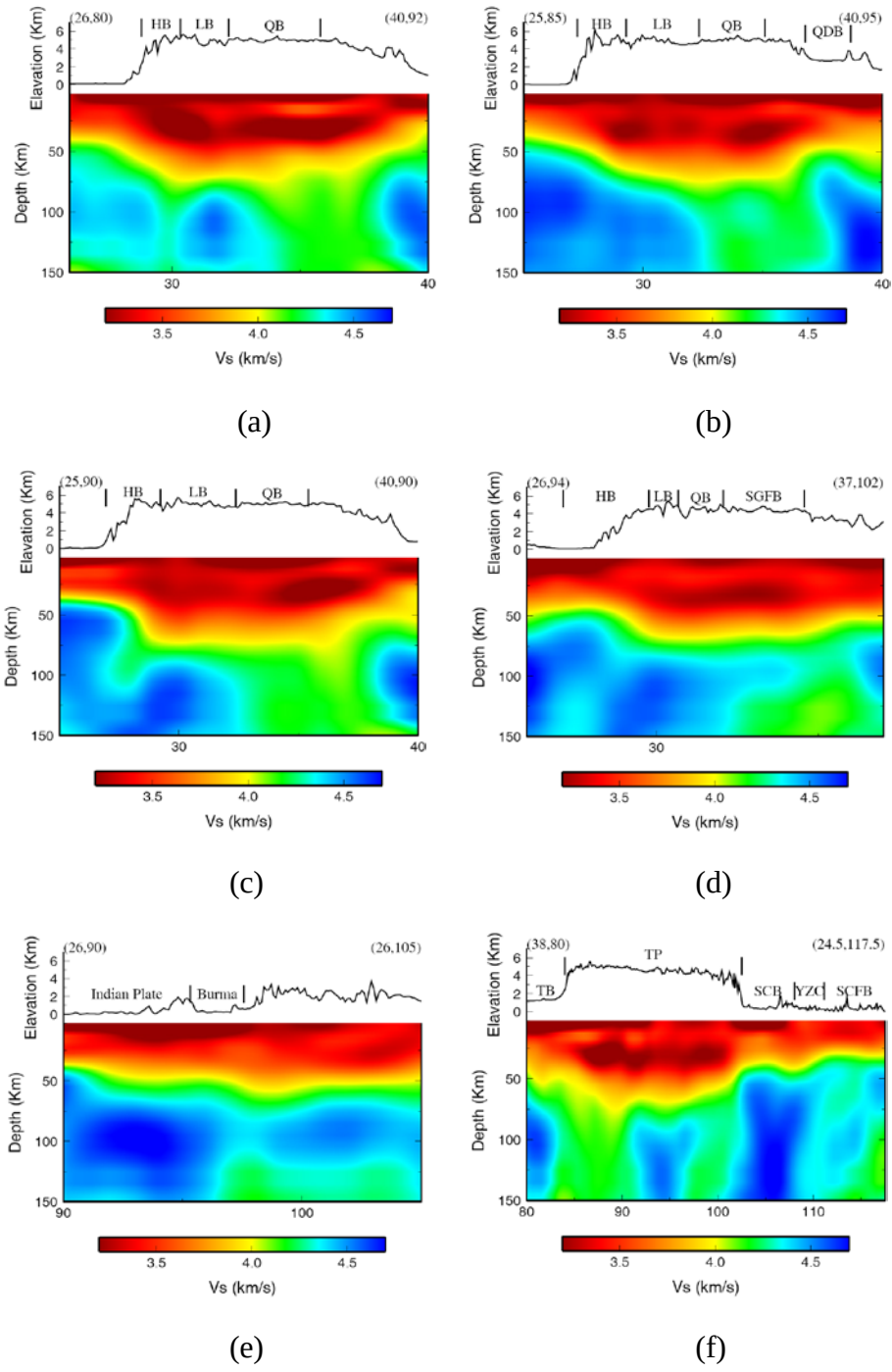
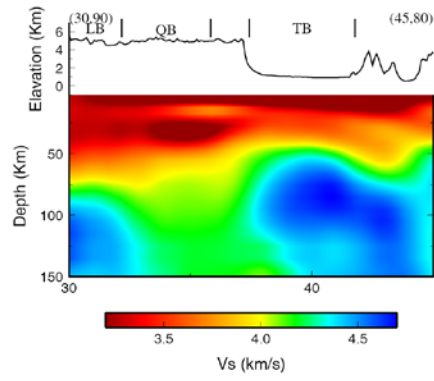
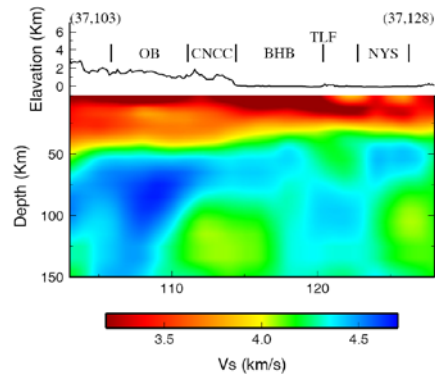


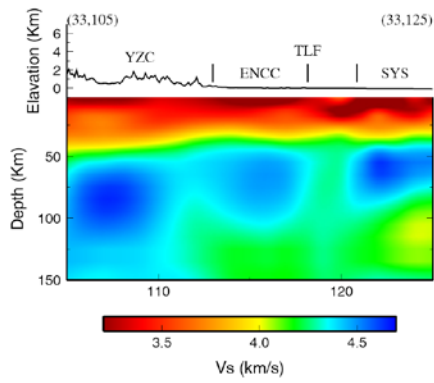
Fig. 2.7



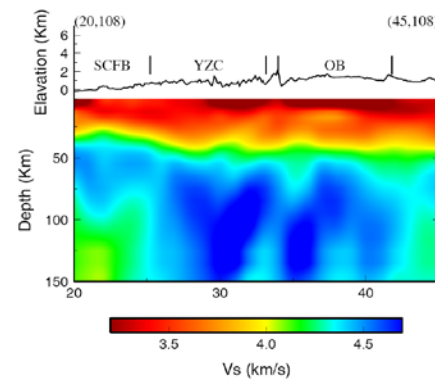
(g)



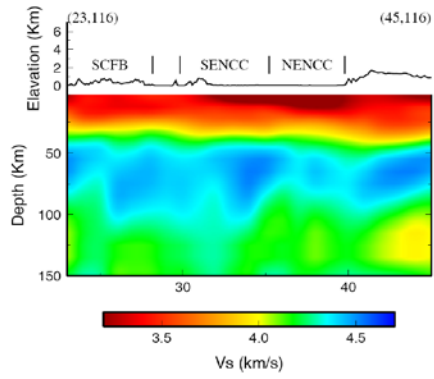
(h)



(i)



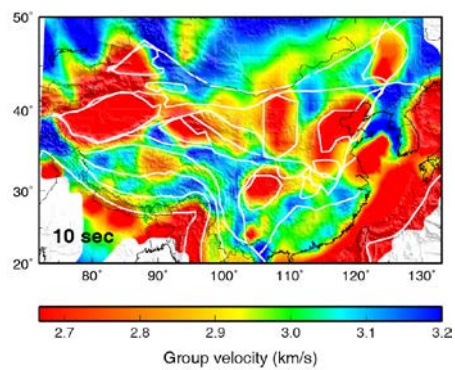
(j)



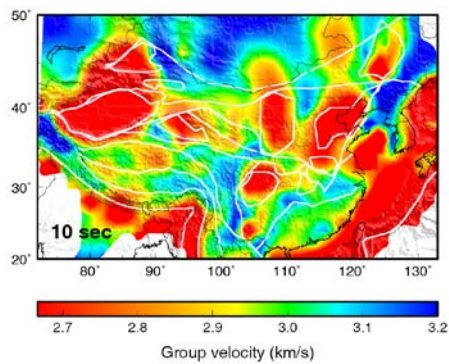
(k)

Fig. 2.7

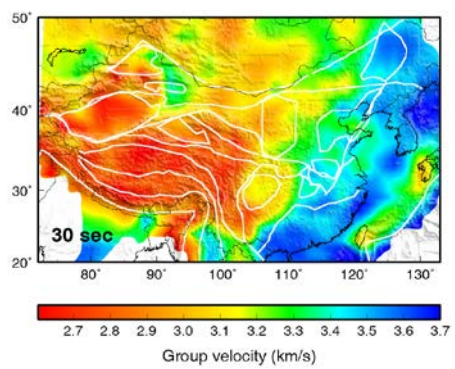
Fig. 2.7. S Velocity cross sections along great circle paths as shown in Fig. 2.6 (a). Prominent low velocity zones are shown in different cross sections across Tibetan Plateau (a-d, f, g). High mantle velocity under Himalaya Block indicates rigid underplating India Plate (a-d). Low mantle velocity in northern Tibetan Plateau suggests India Plate is detached leaving hot and highly deformed Asia mantle materials (a-d, f). Major basins like Sichuan, Ordos and Tarim show thick sediment deposition in shallow depth and deep continent root at greater depth (f-h, j). Velocity and mantle lithosphere contrast between eastern NCC and western NCC is also clearly implicated in (h-k).



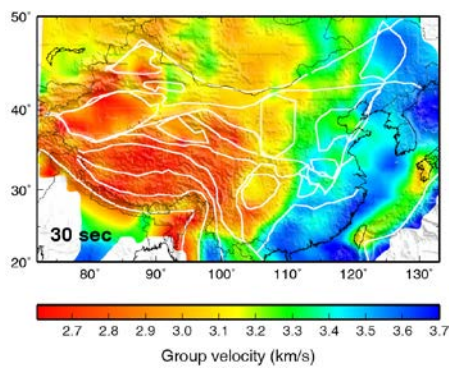
(a)



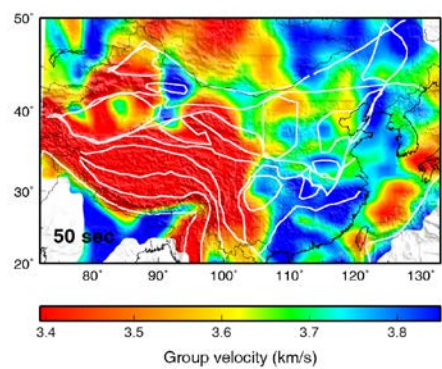
(b)



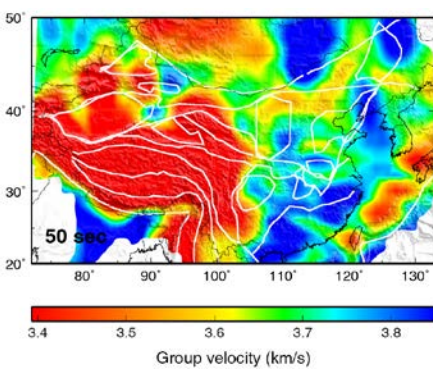
(c)



(d)



(e)



(f)

Fig. 2.8

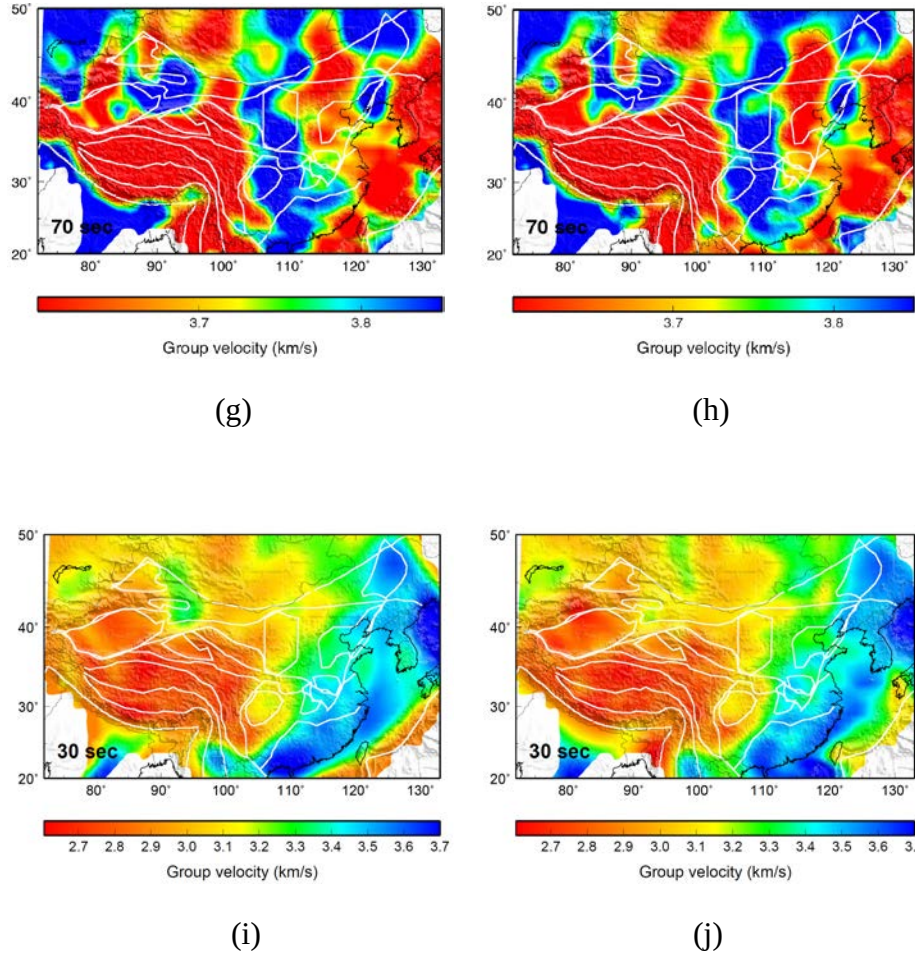


Fig. 2.8. Comparison of group velocity maps using different data. a, c, e, g show the results at 10, 30, 50, 70 s, using combined data while b, d, f, h are the results using earthquake data only at corresponding period. Results from two different data sets are compatible with detailed differences. Major features like basins, Tibetan Plateau, North China Craton are all well resolved. i and j are the group velocity maps at 30 sec using ambient noise data (i) and earthquake data (j), respectively. Note no initial reference model is used in the inversion of i and j.

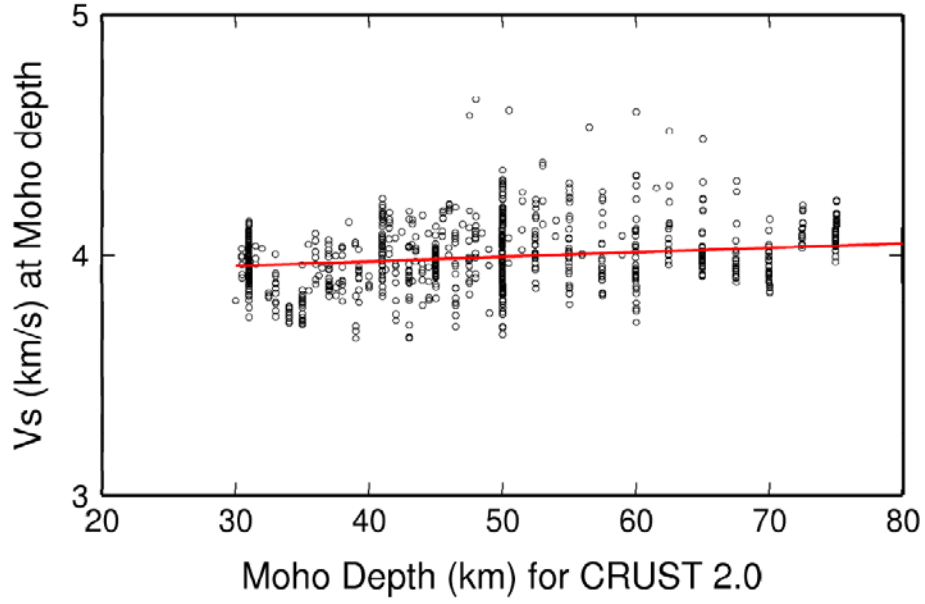


Fig. 2.9. S velocity at the depth of the CRUST 2.0 Moho retrieved from our S model. Each open circle corresponds to each grid in our inversion. The red line is the best fit line, which is used for Moho depth calibration. Note the gentle slope of fitting line, indicating nearly constant S velocity.

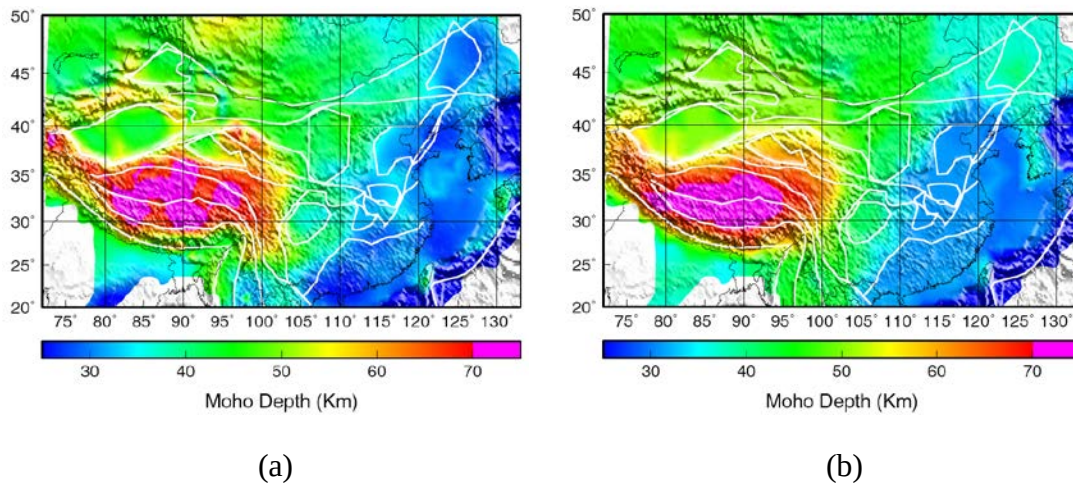


Fig. 2.10. Crustal thickness map derived from S velocity model (a), in comparison with CRUST 2.0 model (b). Note two models are in general agreement with more detailed structure in our model derived from S velocity.

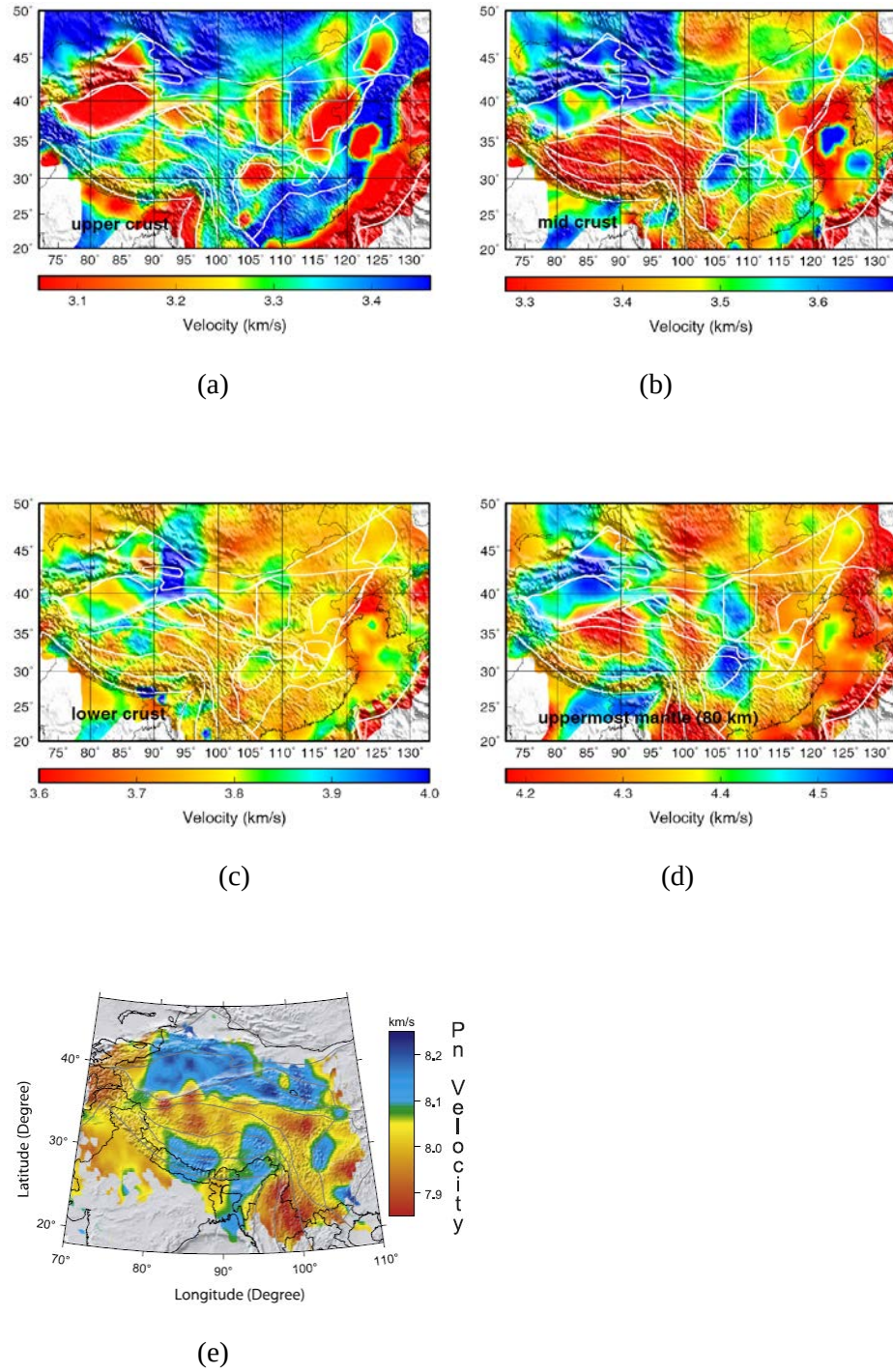


Fig. 2.11. Average S velocity in the upper crust (a), mid crust (b), lower crust (c), and 80 km uppermost mantle (d), in comparison of Pn velocity model in southwestern China (e) from Liang and Song (2006).

CHAPTER 3

JOINT INVERSION OF RECEIVER FUNCTION AND SURFACE WAVE DISPERSION USING NEIGHBORHOOD ALGORITHM: AN APPLICATION TO HI-CLIMB LINEAR ARRAY IN TIBETAN

3.1 Introduction

In seismic inversion, trade-offs between model parameters are well known. To resolve the ambiguity and to improve resolution, a combination of different data sets that have sensitivities to different parameters is required or *a priori* constraints have to be imposed. For example, Receiver functions are an efficient tool to study the structure of the earth's interior including Moho depth (Zhu and Kanamori, 2000), crustal velocity models (Owens et al., 1984), and upper mantle discontinuities (Shen et al., 1998). However, it has long been recognized that teleseismic P receiver functions are sensitive to shear velocity contrast and depth-velocity product, instead of velocity alone. Therefore substantial trade-offs exist between the average wave velocity above the interface and the depth to the interface (Ammon et al., 1990). On the other hand, surface waves are primarily sensitive to the vertical shear-velocity averages while insensitive to velocity discontinuity. Thus, combining these two complementary data by jointly inverting of receiver functions and surface wave dispersions would greatly reduce depth-velocity ambiguity and has now been commonly used to resolve depth resolution of Vs structure (e.g., Julia et al., 2000; Ozalaybey et al., 1997). Similarly, crustal thickness estimated using only the delay time of Moho P-to-S converted phase (Ps) relative to P at the receiver trades off strongly with crustal Vp/Vs ratio. Although it can be remedied by using later converted phases, those multiples usually have less energy, thus are harder to identify and can easily be contaminated by noise. An alternative solution is to add well constraint P wave data, e.g. Pn delay time to decouple the trade-off between crustal thickness and Vp/Vs ratio (Fig. 3.1).

Different research groups have implemented different joint inversion schemes, among which the traditional least square based linearized inversions are widely used (Julia et al., 2000; Ozalaybey et al., 1997). However, due to the nonlinear properties of the inversion problem and the dependence of the starting model based on *a priori* knowledge, linearized inversion may not be the best approach in such inversion. With the improvement of computational power, several

nonlinear searching algorithms have been developed and applied in geophysical inversion problem. Commonly used algorithms include but not limited to the Genetic Algorithm, Simulated Annealing, and Neighborhood Algorithm. The advantages of using searching algorithms include the followings 1) They deal with nonlinear problems easier than traditional linear inversions. 2) They are less likely trapped in local minimums. 3) The location of Moho is naturally defined. 4) Error analysis can be performed in a relatively straightforward way. However, one major drawback of search algorithms is that they are very time consuming. Nevertheless, with the increased computer power and utilization of parallel computation, the computation time becomes less of a concern. In this paper, we adopt the Neighborhood Algorithm (NA) developed by Sambridge (1999a, b). NA is an optimized direct searching algorithm that converges faster than traditional Monte Carlo searches and has been proven successful in various geophysics inversions. It makes use of geometrical constructs known as Voronoi cells in the search and appraisal stages. Recent NA development also implemented Message Passing Interface (MPI) for parallel process (See appendix B) that can greatly reduce the computational time.

The western China, marked by spectacular topographic relief, complex geology, and active deformations, is one of the most active areas of continental earthquakes in the world. Major basins, high plateaus, great mountains all co-exist in this area, which are significantly affected by Indian-Eurasian collision. Geological diversities make western China a hot spot for geology and geophysical studies. Mapping accurate crustal and upper mantle velocity structures will help to answer many exciting geology questions, such as the deformation of Tibetan Plateau during continental collision, the relation between N-S shortening and E-W extension, mid crust channel flow in Tibetan Plateau, subduction of Indian mantle lithosphere, underplating and delamination of Indian crust. With the increase of new permanent and temporary seismic stations deployed in Tibet during recent decades, various seismic studies have been carried out in this region. An integrated study of receiver function inversion, two-station Rayleigh wave phase velocity inversion and teleseismic P-waveform modeling by Kind et al. (1996) using INDEPTH-II seismic stations revealed prominent low velocity zone in the mid-crust north of the Yarlung-Zangbo suture, beneath the southern Lhasa block, while such feature is absent south of the suture under Tethyan Himalaya. Yuan et al. (1997) reported a Moho depth at 70-80 km and a second discontinuity at 50-60 km across entire INDEPTH II profile south and north of Yarlung-Zangbo

suture. They also observed pronounced low-velocity zone north of the Yarlung-Zangbo suture at about 10-20 km depth. An independent receiver function study by Nabelek et al. (2009) using Hi-CLIMB data revealed weak Moho Ps conversion from 31°N to 33°N. Their study also suggested the crust/mantle interface beneath Tibet is anisotropic. The similar Moho variation is also reported by Tseng et al. (2009) using SsPmp phase and by Nowack et al. (2010) using Gaussian beam migration. Hetenyi et al. (2006) estimated the effective elastic thickness (EET) of Indian Plate from receiver function images, gravity anomalies and thermomechanical modeling. Their results suggest the EET of Indian Plate drops from ~60-70 km under Indian Continent to ~20-30 km north of Main Frontal Thrust (MFT) due to the loss of the crustal strength. They also suggest a strong mantle in the study region between Ganges basin and Central Tibet in order to be consistent with modeled flexural geometry. Wittlinger et al. (2009) reported the detection of eclogite layer in the lower crust under Southern Lhasa block from grid search stacking of receiver functions. Hung et al. (2010) presented a first multi-scale finite-frequency P and S velocity model in Southern and Central Tibetan Plateau, revealing sub-horizontally advancing Indian lithospheric mantle beyond Bangong-Nujiang Suture as well as regions of low velocity structures in the crust that correlates with low electric resistivity study (Unsworth et al. 2005). Sun et al. (2010) performed Rayleigh wave tomography in China region using ambient noise correlation data. They reported a wide spread low velocity zone in Tibetan Plateau in the mid crust, similar to those reported by receiver function studies. By comparing the Rayleigh wave and Love wave dispersion, Shapiro et al. (2004) predicted radial anisotropy with $V_{SH} > V_{SV}$ in the mid to lower crust. The anisotropy is the largest in western Tibet, corresponding to the thinning of mid crust. Although some first order common features are shared among these aforementioned studies, discrepancies and controversies still exist, e.g. channel flow vs. isolated low velocity zone, strong mantle vs. weak mantle. In this paper, we jointly invert surface wave group and phase velocity and receiver functions from Ganges basin in Indian Plate to Central Tibet using seismic data recorded by Hi-CLIMB experiment stations. The near north-south dense linear array provides ideal geometry to construct high resolution 2D shear velocity profile.

3.2 Data and Method

For Rayleigh wave dispersion, we requested all the long period continuous vertical component data for 150 temporary stations (H0010 to H1630) from Sep. 2002 to Aug. 2005. These stations

consist of a dense linear array that extends from the Ganga foreland, across the Higher Himalaya, then continuing across the Tsangpo suture into central Tibet (Fig. 3.2). Average spacing between two adjacent stations is ~ 5 km. (Nebelek, et al., Project summary). Ambient noise data are processed following the detailed description by Bensen et al. (2007). We first synchronize the time of one-day record for each station and bandpass filter the data from 3 to 120 s. Temporal and spectrum whitening is performed after filtering process to remove coherent energetic signal. We compute cross correlation of possible station pairs on daily bases. The calculated daily cross correlations for the same pair are further stacked to enhance signal-to-noise ratio (SNR). The symmetric part of Empirical Green's Function (EGF) is obtained by averaging the positive and negative lags. The SNR for the symmetric EGFs is estimated and those with $\text{SNR} > 10$ are selected to measure the Rayleigh wave group and phase velocities from 4 to 45 s using traditional frequency time analysis (FTAN) (Levshin and Ritzwoller, 2001). We also require the inter-station distance should be larger than 3 wavelengths to satisfy far field approximation.

The profile is parameterized into grid with spacing of 0.1 degree in latitude. 1D linear inversion is performed for each period. During the 1D inversion, we use group and phase velocity from China entire surface wave inversion (Xu and Song, in preparation) as an initial model. The group and phase velocities at each node are obtain, which is compiled further as the input of the joint inversion.

The receiver function (RF) data for the corresponding stations are processed and provided by Lupei Zhu in St. Louis University. RFs for each station are stacked according to ray parameters. For each station, we pick four stacked RFs with clear Ps converted phase and different ray parameters for the inversion. The number of events for each stacked RF varies from 3 to over 50, depending on the distribution of the events and the quality of the data.

Our joint inversion is a two-step scheme. In the first step, we invert only the dispersion data to obtain S velocity profile at each station. We use the result from the first step as the initial model for the searches of the final velocity model using both dispersion and RF data. It has been proven that using a good initial model will help the search converge faster. We use cubic spline in depth for model parameterization. The advantage of using splines is that they yield naturally smooth model and yet reduce the number of parameters for searching. For dispersion inversion in the

first step, we adopt 12 uniformly distributed spline nodes from the surface down to 150 km. We seek a solution that satisfies the following minimizer:

$$\min (\|\mathbf{D}^{\text{obs}} - \mathbf{D}^{\text{pred}}\| + \lambda \|\mathbf{m} - \mathbf{m}_0\| + \varphi \|\mathbf{Lm}\|) \quad (1)$$

In the above formula, the first term is data misfit, where $\mathbf{D}^{\text{obs}}$ is the data we are trying to fit, and $\mathbf{D}^{\text{pred}}$ is the data predictions from the random model generated by NA algorithm. The second term is the model constraint. Since the periods used in the inversion only extend to 45 s, which has poor constraint on the deeper part of the structure, we introduce model constraint for the depth from 95 km down to the 150 km using the S model from China country inversion, which includes group velocity period up to 120 s. The third term is the smoothing constraint. $\mathbf{L}$ is the Laplacian smoothing operator (Lees and Crosson, 1989), which requires second derivative of the model to be zero. We seek the solution to minimize the L1 norm of above formula instead of traditional L2 norm. The choice of L1 norm reduces the risk that the model is biased by extreme values due to error in the data. The damping parameter λ determines the influence of the initial model and smoothing parameter φ determines the smoothness of the model. A larger λ value or φ value makes the solution more stable or smoother but increases the data misfit. We experiment different values and used $\lambda=1$ and $\varphi=0.3$ in our searches. We convert the spline model into a layered velocity model to calculate predicted Rayleigh group and phase velocity. Velocities in the middle of each layer are calculated from spline coefficients to represent the velocity of that layer. We divide the whole 150 km into 30 layers with 5 km thickness for each layer. Velocities within each layer are constant. We fix the V_p/V_s ratio for all the layers to 1.75. Density is calculated by the Nafe-Drake relationship (1963). The forward calculation subroutine is adopted from Robert Herrmann's program (1991).

The parameterization for the second step joint inversion is different from the first step. We use two sets of spline to represent velocity model. The last node of the first set of splines has the same value as the first node of the second set of spline so that the model is continuous throughout the depth. The node spacing for each set of spline is not uniform; instead, it is stretched over depth. The nodes are closer to each other near the top and farther apart at the bottom of the splines. We use the following formula to map the depth into normalized spline space from 0 to 1 (See Appendix B).

$$h = Dx^t + h_0 \quad (2),$$

where h is the depth from the top of the first node in the spline; D is the total thickness; h_0 is the depth of the first node; x is the coordinate in the normalized spline space between 0 and 1. t denotes stretching parameter. The larger the t is, the more the nodes are stretched. If t equals 0 than formula 2 is simplified to traditional uniform spline case. In this paper we find $t=2$ is an appropriate number. This set up is efficient modeling the structure at shallow depth and near Moho discontinuity region by giving more freedom to the model where it is better constraint by short period dispersion data and receiver function, respectively. It also yields smoother model at the lower crust and deeper mantle, which are less constraint by the data. We use 12 nodes for the first set of splines in the crust and 10 nodes for the second set of splines in the mantle. We also allow the depth of the connecting node between the first set of spline and second set of spline to vary so that the Moho depth can be adjusted according to the P_s arrival time. For a velocity model with 70 km crustal thickness, the node spacing between the first node and second node in the mantle is ~ 0.99 km and the spacing between the second and third node is ~ 2.96 km, enough to model sharp velocity jump, thus mimic the Moho discontinuity. This parameterization also works well in the situation where Moho is not sharp but presents as transition zone. We convert the spline model into a layered model as in the first step. We use 2 km layer thickness throughout the depth range of the first spline set in the crust. In the mantle, we use fifteen 2 km layers for the top 30 km as transition zone and followed by 5 km layer to the bottom at 150 km. We seek solution that satisfies the L1 norm of the following minimizer:

$$\min(w_1 \| \mathbf{D}_{\text{disp}}^{\text{obs}} - \mathbf{D}_{\text{disp}}^{\text{pred}} \| + w_2 \| \mathbf{D}_{\text{rfs}}^{\text{obs}} - \mathbf{D}_{\text{rfs}}^{\text{pred}} \| + \lambda \| \mathbf{m} - \mathbf{m}_0 \| + \varphi \| \mathbf{Lm} \|) \quad (3),$$

where the first term is dispersion misfit, the second term is RFs misfit and the third and last terms are the same as formula (1). The w_1 and w_2 are the relative weighting for dispersion and RFs misfit, respectively. We experimented different w_1 and w_2 values and chose $w_1=1$ and $w_2=0.25$ in our model searches.

Every time when NA generates a random model, we compare the model with initial model from dispersion inversion for each layer except those 15 layers in the transition zone. If the velocity difference in any given layer is larger than 0.5 km/s, this model is discarded by setting a large misfit value without going through the forward calculation and new model is generated. We also

require the velocity to be increase monotonically from the last two nodes in the crust to first two nodes in the mantle. This model screening process can reduce computational time by confining the model search in the model space where the likelihood is the largest. In addition to model screen, we further introduced Ps delay time to guide the search (See Appendix B). Fig. 3.3 shows the delay time measurements for each station at ray parameter of 0.06 s/km. Red curve is the spline fit of the Ps delay data. Two complicated zones from $\sim 27.5^{\circ}\text{N}$ to $\sim 29^{\circ}\text{N}$ and from $\sim 31.5^{\circ}\text{N}$ to $\sim 33^{\circ}\text{N}$ are presented by scattered Ps delay time. We calculate the departure of the Ps delay data to its corresponding fit for each station. If the difference is larger than 1.5 s, then we use the predicted Ps delay time from the best fit curve as the input, otherwise we use the original Ps delay time as the input. In the search, we calculate the velocity gradient at each layer and pick the one with largest gradient as the bottom of the crust. In most cases, it is the last layer in the first set of spline. We calculate the predicted Ps delay time from the random model generated by NA and compare with Ps delay data, if the difference is larger than 1 s, then the model is discarded. Our experience suggests by introducing the Ps delay time, the model converges faster than without Ps delay time data. In the joint inversion, without Pn delay time data, we also fix the V_p/V_s ratio to 1.75 throughout the depth range. Density of each layer is calculated using the same Nafe-Drake relationship as described above. Because of the lack of coherent crust multiple signals, only the first 15 s of receiver functions are included in inversion. For each station, we select 4 receiver functions with different ray parameters for inversion. Subroutines of dispersion and receiver function forward calculation are adopted from Robert Herrmann's computer programs. The NA inversion is following the description by Sambridge (1999). We iterate 800 times to obtain the final model. Initially, we generate 200 random models to start with. The model space is divided into Voronoi cells based on the generated random models so that each model lies in one Voronoi cell. 5 Voronoi cells with highest ranks (smallest misfits) are resampled and a total of 100 new models are generated within these 5 cells. The total models generated for each station is 80200.

3.3 Results

Fig. 3.4 (a) shows color image of the shear velocity profile along Hi-CLIMB linear array from joint inversion. A pronounced low velocity zone at depth of around 25 km in the mid-crust exists in most part of the Tibetan Plateau. However, the thickness and velocity reduction varies at

different locations. Under the Himalaya block near Yarlung-Zangbo suture, the low velocity zone can be over 25 km thick with velocity reduction of more than 10% while under Lhasa block both thickness and velocity reduction decrease significantly while the amplitude of low velocity picks up again under Qiangtang block. However, it seems that the low velocity feature is not continuous across Bangong-Nujiang suture. The low velocity feature near Yarlung-Zangbo suture is also found by multi-scale finite-frequency body wave tomography (Hung et al., 2010) and correlates with low electric resistivity (Unsworth et al., 2005), suggesting the presence of fluid or partial melt. Crustal thickness varies from about 50 km in the southern part of the profile under Ganga foreland in Indian plate to about 75 km under south Lhasa block and gradually decreases to less than 65 km under Qiangtang block. This trend agrees well with the Moho depth determined by Nabelek et al. (2009) using receiver function alone (white line in Fig. 3.4), by Tseng et al. (2009) using SsPmp converted phase, and by Nowack et al. (2010) using Gaussian beam migration. The location of Moho discontinuity under each individual station is shown in crosses or circles, where crosses indicate the stations with clear shear velocity jump while circles mark the place where the shear velocity jump is smaller than 0.15 km/s across the Moho discontinuity. Most circles and large Moho variations locate from 27.5°N to 29°N, where the topology changes drastically from the low land to the highly elevated plateau over a relatively short distance and from 31.5° N to 33°N in northern Lhasa block near Bangong-Nujiang suture. The complex Moho discontinuity coincides with scattered Ps delay time measurements (Fig. 3.3). Particularly, the northern segment of blurred Moho is consistent with the region where Moho Ps conversion is weak (Nabelek, et al., 2009, white dashed line in Fig. 3.4a) and a disrupted Moho is identified from Gaussian beam migration (Nowack, et al., 2010). The uppermost mantle shear velocity is slower north of 31° N, consistent with S velocity profile from finite-frequency tomography (Hung, et al., 2010).

Fig. 3.4 (b) shows the velocity profile under stations. One velocity profile is chosen from every three stations. Moho depth variations and low velocity features similar to Fig. 3.4 (a) are also presented here. It is noticeable that there seems to have a depth offset between the low velocities south and north of Bangong-Nujiang suture. The amplitude of low velocity zone is small and occasionally not clear under Lhasa block.

Fig. 3.5 shows the fit of dispersion and receiver functions as well as 1D velocity profile for example stations from south to north. Both waveforms of receiver functions and the shape of dispersions, particularly at short periods are different from southern station to northern station. There is no low velocity zone under station H0330 in the south. The Moho velocity jump is very clear at ~52 km. Station H0720 locates in Himalaya block, south of Yarlung-zanbo suture. A broad low velocity zone appears in the mid-crust from 14 km to 44 km with a maximum velocity reduction of 11%. A clear Moho velocity jump occurs at 68 km. Station H1220 locates in the southern Lhasa block. The 1D shear velocity profile shows a low velocity structure near 25 km. The Moho depth reaches almost the thickest in the profile at ~75 km. Velocity structures between 31.5° N and 33°N are complicated and seem to show no consistency between different stations. H1310 and H1320 are two stations next to each other. Weaker but still distinguishable mid-crust low velocity zones are imaged under both stations. However, H1310 exhibits a much sharper Moho discontinuity than H1320, where the Ps converted phase is weak and inconsistent. H1350 is the station in the same region. The crust velocity under H1350 does not resemble any of previous models: The Moho velocity jump is large while the low velocity zone in the mid crust is almost absent. Further in the north, H1620 shows a sharp but shallower Moho at 63 km, while the mid crust low velocity zone is weak and broad, extending to lower crust. A typical character of the low velocity zones under these stations is an increase of velocity from surface to about 10 km depth, followed by velocity decrease to its minimum at around 20 km then gradual increase again. Synthetic tests suggest that this feature is mostly controlled by group dispersion in short period range (less than 15 s), where the group velocity increases from 4 s to its maximum at around 10 s and then gradually decreases. The stronger the low velocity zone is the more pronounce the feature is (comparing Fig. 3.5b with Fig. 3.5f).

To examine the robustness of our final models, we plot the sampled models in the search (see examples of representative stations in Fig. 3.6). The blue thick line is our best fit model. The light greys in the background in each panel indicate all the ranges of all the 80200 models in the search. The dark greys are the best 200 models with smallest misfits. They have similar misfit value, thus we consider them in the same category. It is noticeable that the NA sampler densely sampled the whole model space as we specified in the search ranges. On the other hand, the best 200 models with smallest residuals all fall in a very narrow range, indicating the converging best models converges and falls into the same category of models. For example, the Moho depth

variations are typically less than 2 km between the best 200 models while velocity variations are less than 0.1 km/s at the same depth between these 200 best models. Therefore we are confident that NA is able to find global minimum in the search ranges to equation (3). The variations of the best fitting models provide a good estimate of the variation of the model parameters. One noticeable exception is station H1620 (Fig. 3.6g), where the final model and other best fitting models differ by up to 0.2 km/s at depths 70-100 km in the mantle. The structures at these depths are constrained by dispersion data only at longer periods.

To test model sensitivity to data errors, we performed a series of random error tests. We choose station H1220 as an example, which has clear Ps converted phase, a pronounced low velocity zone, and a sharp Moho. In the test, we start with the best fit model. We allow each parameter of the model, i.e. spline coefficients and Moho depth, to perturb randomly over a large range of up to $\pm 20\%$ with a uniform distribution. Each time when a new model is generated, we calculate the predicted receiver functions and surface wave dispersion and compare the predictions with the data. We run the tests for 80000 times, which assure enough sampling of the possible model space based on the smoothness assumption of cubic splines. We select the models according to certain criteria. The selection of model selection criteria requires careful consideration. In our tests, we use Ps delay time, the dispersion group and phase velocity values at a range of periods, and the average amplitudes of P and Ps at four receiver functions of different ray parameters. These parameters are the most robust features of the data that we try to fit in the inversion. The departures of model predictions from the observations for these parameters are also straightforward to quantify.

Although we are not able to directly assess the error in dispersion and stacked receiver function, Zheng et al. (2008) obtained one standard deviation of dispersion measurement error to be less than 2% for all the periods for the pairs they examined, using stacks of Green's function of different time periods. We also roughly estimate the one standard deviation of Ps-P time to be 0.19 s, using the measurements from receiver function data in the region between 30.1°N to 30.9 °N, where the Moho is more or less flat. Here we present two results with large data errors. Fig. 3.7(a, left) shows the selected models that have Ps-P delay time differing from the data by 0.2 s; the difference of group and phase velocity at each period between the data and predictions

is less than 6% and the differences of maximum P and Ps amplitudes between the data and predictions are less than 10%. It appears that the model is less scattered in the crust. Given the scatters, it seems that the low velocity features in the crust is still visible, however the strength may vary between different models. Large variation occurs in the mantle velocity due to poor constraint from the data and Moho depth because of depth-velocity trade-off and the allowance of Ps-P delay time to vary. Fig. 3.7(a) right panels show the predictions of receiver functions and surface wave dispersions from the selected models. Those predictions cover a large range that represents the possible error in the data. Fig. 3.7(b, left) shows the selected model similar to Fig. 3.7(a, left) but we allow Ps-P delay time difference between the data and predictions to be as large as 0.3 s and the difference of group and phase velocity, and the difference of maximum P and Ps amplitude between the data and predictions to be up to 10% and 20%, respectively. This set of criteria assumes larger error in the data. More models are selected due to the extremely loose criteria. The predictions from selected models cover a broad range. Under such conditions, the low velocity zone in the crust may not be resolvable.

3.4 Discussion

Our joint inversion image shows the mid crust low velocity structure starts at the edge of Tibetan Plateau and appears in most area along the profile, in favor of mid-crust low velocity zone model as reported by the receiver function study of Kind et al. (1996) and the surface wave tomography of Sun et al. (2010). The low velocity feature suggests that the mid crust material may be partially molten. However, the low velocity structures seem to form patches instead of continuous flow channels as suggested by Clark and Royden (2000) and are discontinuous near Bangong-Nujiang suture. Also the thickness and amplitude of the low velocity varies at different location. Our disconnected low velocity zone model is in agreement with the results by Hung et al. (2010), suggesting the popular channel flow model may need to be revisited. The model derived by Nabelek et al. (2009) using Hi-CLIMB data, on the other hand, is against the existence of mid crust low velocity zone, based on the absence of coherent low velocity zone signal and normal Vp/Vs ratio. However, our experiments suggest the low velocity feature near 25 km is required in order to fit the shape of short period dispersion data. We also notice that the signals in receiver functions between the first P arrival and Ps converted phase are very complicated. Therefore the signal from mid crust low velocity zone may be overwhelmed by the

complicated multiples/arrivals within the crust and/or high noise level. Because of the limitation in receiver function, without knowing precisely the P wave velocity structure, only average crust Vp/Vs ratio can be estimated, even including crust multiples like PpS and PsS. Given the thickened crust of about 70 km, a localized magma and/or fluid may not be sufficient to change the average crust Vp/Vs ratio significantly (Kind et al., 2002).

Our joint inversion result also suggests crust is thicker under Lhasa block while thinner under Qiangtang block. The northward thinning of crust under Tibetan Plateau is in consistence with many previous receiver function results using the same Hi-CLIMB stations (Nabelek et al., 2009, Tseng, et al., 2009, Nowack, et al., 2010) and similar to the results using INDEPTH stations in the east of our profile (Kind et al. 2002, Shi et al. 2004). However, the topology is more or less flat across the whole Tibetan Plateau. Therefore the varying of crustal thickness indicates Airy isostasy is not reached throughout the whole Plateau. In fact, the study by Tseng et al. (2009) suggests southern portion of the Lhasa block is close to Airy isostasy while over 10 km crust material under Qiangtang block is missing. It is believed that the thinner crust under northern Tibet is compensated by hot upper mantle material below lithosphere (Tseng et al., 2009, Owens and Zandt, 1997). This hypothesis can be supported by the most recent surface wave tomography by Sun et al. (2010) and Xu et al. (in preparation), which suggest the upper mantle S velocity is exceptionally slow under Qiangtang block.

Nabelek et al. (2009) proposed a zone located north of 31°N and south of the Bangong-Nujiang Suture as the boundary between two lithospheric-scale blocks: To the south, the Tibetan plate is underplated by Indian plate, and to the north is Eurasian plate. This zone coincides with the zone of complicated Moho structure in the northern part of the profile where the Ps arrivals are scattered over short distance range. A similar result is obtained from Gaussian beam migration (Nowack et al., 2010). Therefore, the discontinuities identified in our joint inversion as “Moho” in this region may be intra-crustal interfaces instead. Wittlinger et al. (2009) discovered a high velocity low Vp/Vs ratio eclogite layer under Lhasa block. The large intra-crust impedance may be related to the mafic and ultramafic materials as indicated by Nowack et al (2010). Moreover, our joint inversion result suggests the crustal structures in this area are different from other region (Fig. 3.5): The mid crust low velocity seems to be discontinued in this zone, indicating little partial melt/fluid in this zone. Therefore, it is reasonable to believe the complexity of this

zone is a result of mechanical deformation instead of magmatic activities (Tseng et al 2009). It is noticeable that the uppermost mantle velocity under is slower north of 31°N . If this is the region where India lithosphere meets with Eurasian lithosphere, the low velocity in the uppermost mantle could be a result of strong deformation and downwelling mantle flow. On the other hand, finite-frequency tomography by Hung et al. revealed a gentle northward dipping Indian mantle lithosphere that extends beyond Bangong-Nujiang suture and no evidence of down-welling between depths of 100 to 400 km. Their results are in contradiction with the tomography result by Tilmann et al. (2003) using INDEPTH data, in which a seismic image of downwelling Indian Mantle lithosphere is reported under central Tibet. However, the debate is beyond our resolvable depth limit. A high resolution image that can extend to greater depth is required to further investigate this controversy. In addition to the transition zone in the northern part of profile, we also identified a second complicated region in the south from 27.5°N to 29°N . The receiver function image by Nabelek et al. (2009) indicates the Main Himalayan thrust fault extends from Nepal on the surface into mid crust under southern Lhasa block. The complexity of Ps time may be related to the underthrusting Indian plate and/or the multiples/arrivals of the converted phase at velocity interface inside the crust and cannot be well modeled due to smoothing model from spline parameterization.

3.5 Conclusion

We have developed a joint inversion scheme to invert surface wave dispersion and receiver function data using neighborhood searching algorithm. The parameterization of two sets of stretched splines serves well for our purpose, with the ability of generating high resolution velocity structure in upper crust and Moho transition zone where constraints from dispersion and receiver function data are better, while yield smooth model in the lower crust and deeper mantle where the model is less constraint by the data. With the initial model from dispersion inversion and model screen, we are able to speed up the searches significantly.

Joint inversion generates sharper Moho image, generally in agree with previous receiver function using the same seismic stations (Nabelek et al. 2009, Tseng et al., 2009, Nowack et al., 2010). The Indian crust is observed to subduct and underplate the Tibetan crust to between Yalu-zangbu suture and the Banggong-Nujiang suture, at the latitude of about 31.5° . The crust thickens from

about 50 km below the Indian foreland to over 75 km south of Bangong-Nujiang suture and turns shallower at around 65 km under Qiangtang block. We identified two zones with complicated Moho structure at 27.5°N to 29°N and at 31.5°N to 33°N. The location of transition zone in the northern part of the profile is consistent with the observations by other studies, suggesting the crust and uppermost mantle is highly deformed. The southern region, although not reported by Nabelek et al., is consistent with scattered Ps arrivals and may be due to complicated crustal structures from the underthrusting of Indian plate.

A low velocity zone is present in the mid-crust in most parts under Tibetan Plateau with up to more than 10% velocity reduction. However the amplitude and thickness of low velocity zone varies and seems to be discontinuous near the Bangong-Nujiang suture. This observation is compatible with the results from surface wave tomography. The low velocity structure seems to form patches instead of channels as modeled by Clark and Royden (2000).

Adding well constraint P wave data in joint inversion is essential to decouple the trade-off between Vp/Vs and Moho depth. An accurate Vp/Vs ratio also helps to determine the physical properties, e.g. partial melt, of crustal material.

3.6 References

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3.7 Figures

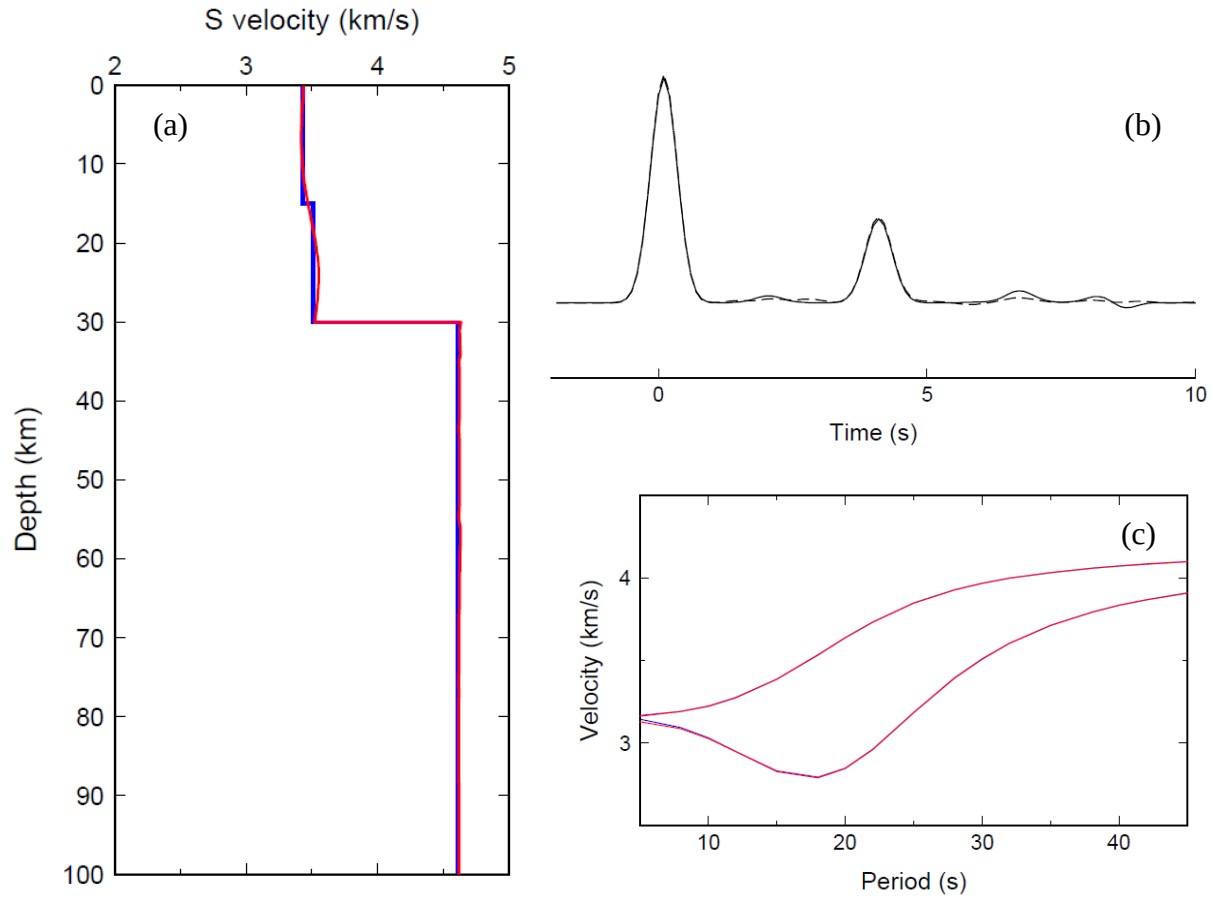


Fig. 3.1. Synthetic test of joint inversion including receiver function, surface wave dispersion and Pn delay time data using Neighborhood Algorithm. Input velocity model is shown in blue curve in (a). Input model consists of three layers. Moho discontinuity locates at 30 km depth. The V_p/V_s ratios in the crust layers are 1.75 and 1.85, respectively. Two sets of splines are used in inversion. Each set consists of 5 evenly spaced nodes. Output velocity model is plotted in red in (a). Note the velocity jump in the crust is smoother than input model due to the natural smoothness of spline function. Average V_p/V_s ratio of 1.80 in the crust is obtained during searches when Pn delay time data is included. The input (solid line) and output (dashed line) of receiver function waveforms are shown in (b). Only the first 10 second of receiver function is used in inversion. Rayleigh wave group and phase dispersion data (blue) from 5 s to 45 s are used in inversion (c). The corresponding dispersion predictions from output model are shown in red.

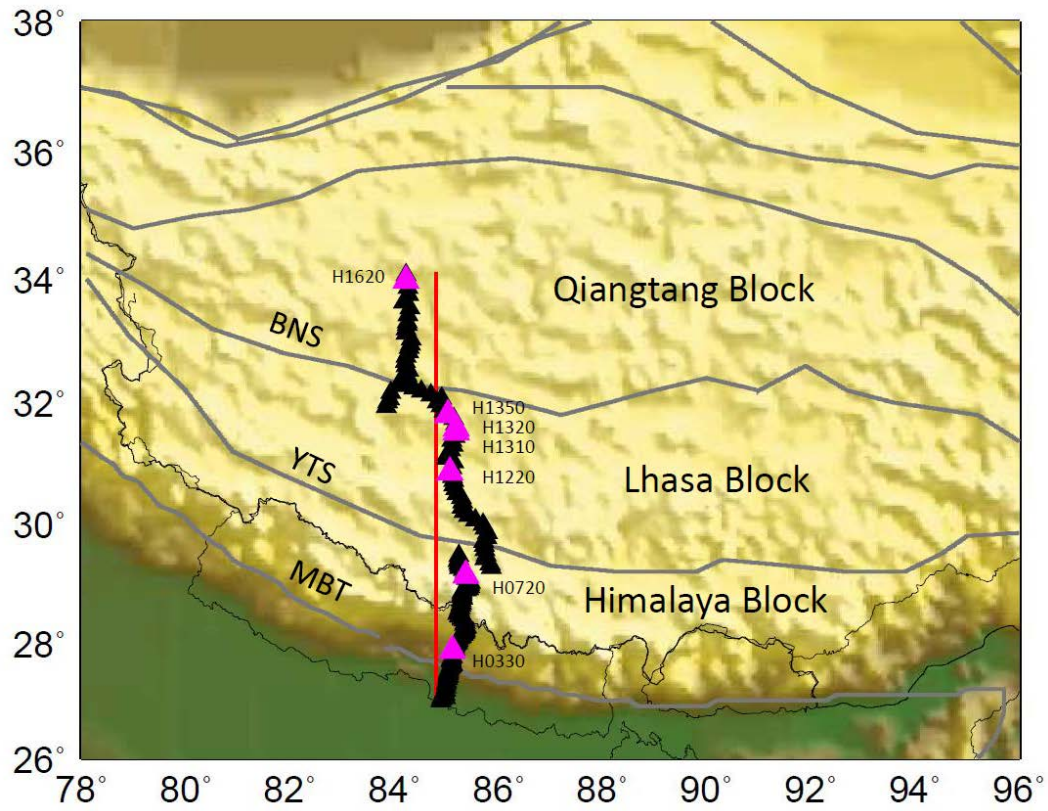


Fig. 3.2. Major tectonic blocks of Tibetan plateau and station distribution of Hi-CLIMB linear array used in joint inversion. Black triangles indicate the position of seismic stations. Purple large triangles are the example stations (with the name starting with H) in Fig. 3.5. The red line is the projection line of the profile as shown in Fig. 3.4. Grey thick lines mark the boundary of major blocks, including Main Boundary Thrust (MBT), Yarlung-Tsangbu Suture (YTS), and Bangong-Nujiang Suture (BNS).

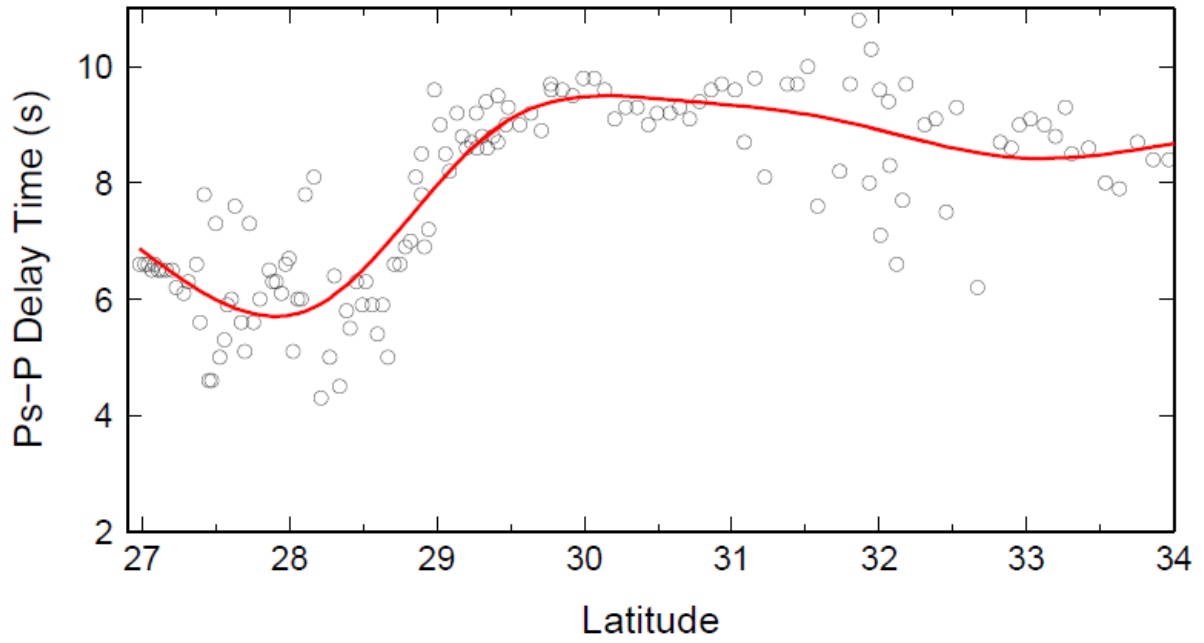
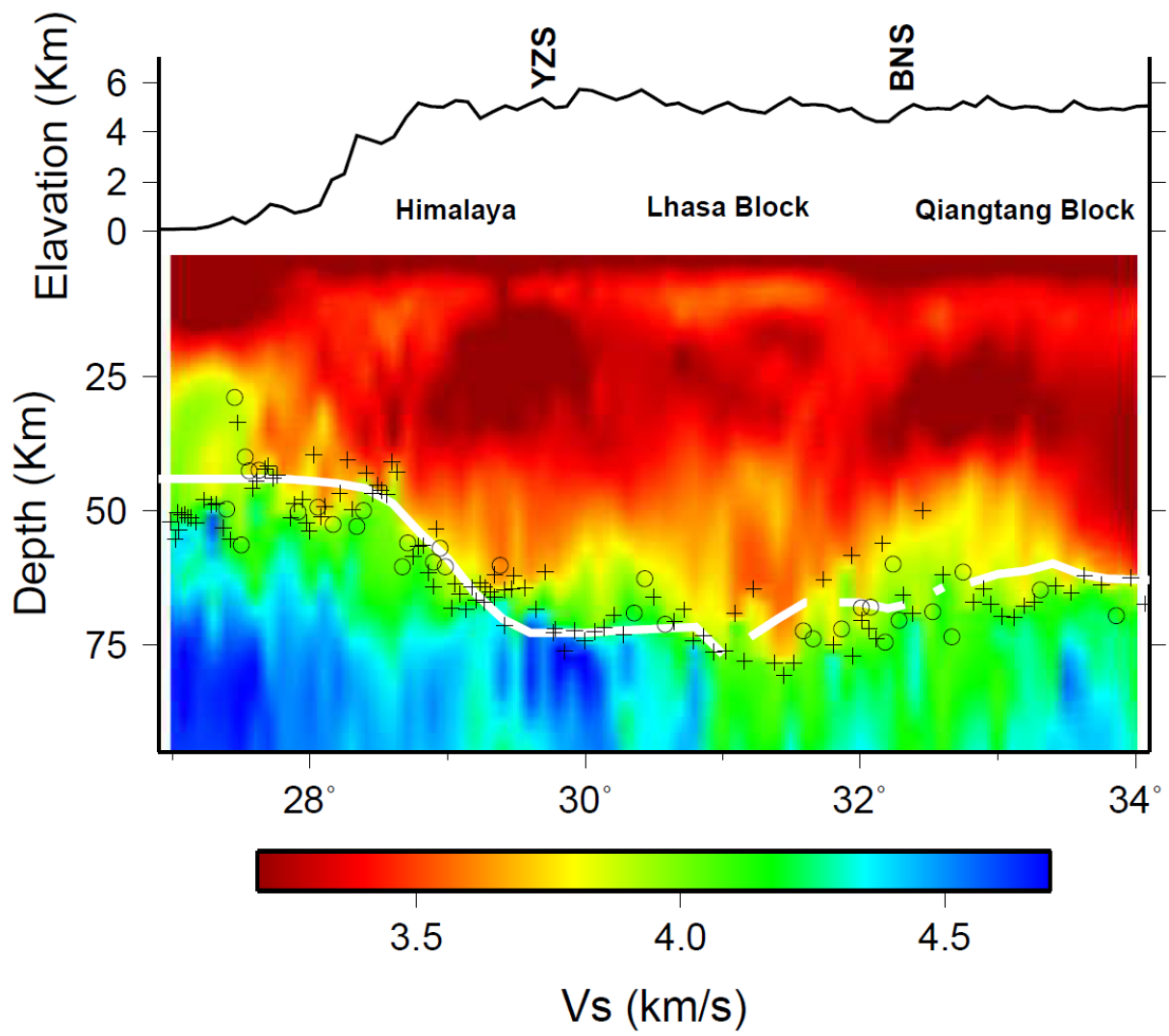
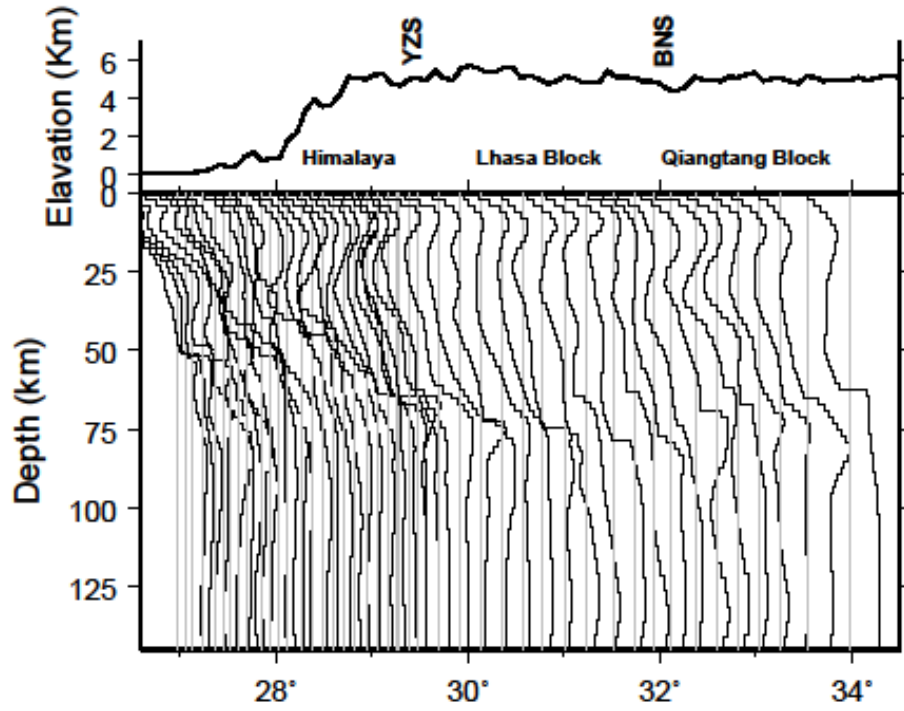


Fig.3.3. Ps-P delay time measurements from receiver function data. Circles indicate the delay time for each Hi-CLIMB stations. Receiver functions for the same seismic station are corrected to ray parameter of 0.06 s/km before stacking. Ps delay time is measured from stacked receiver function. Red curve is the least square fit of all Ps delay time data using spline function with 7 evenly distributed nodes. Note the data is scattered around 27.5°N to 29°N and 31.5°N to 33°N.



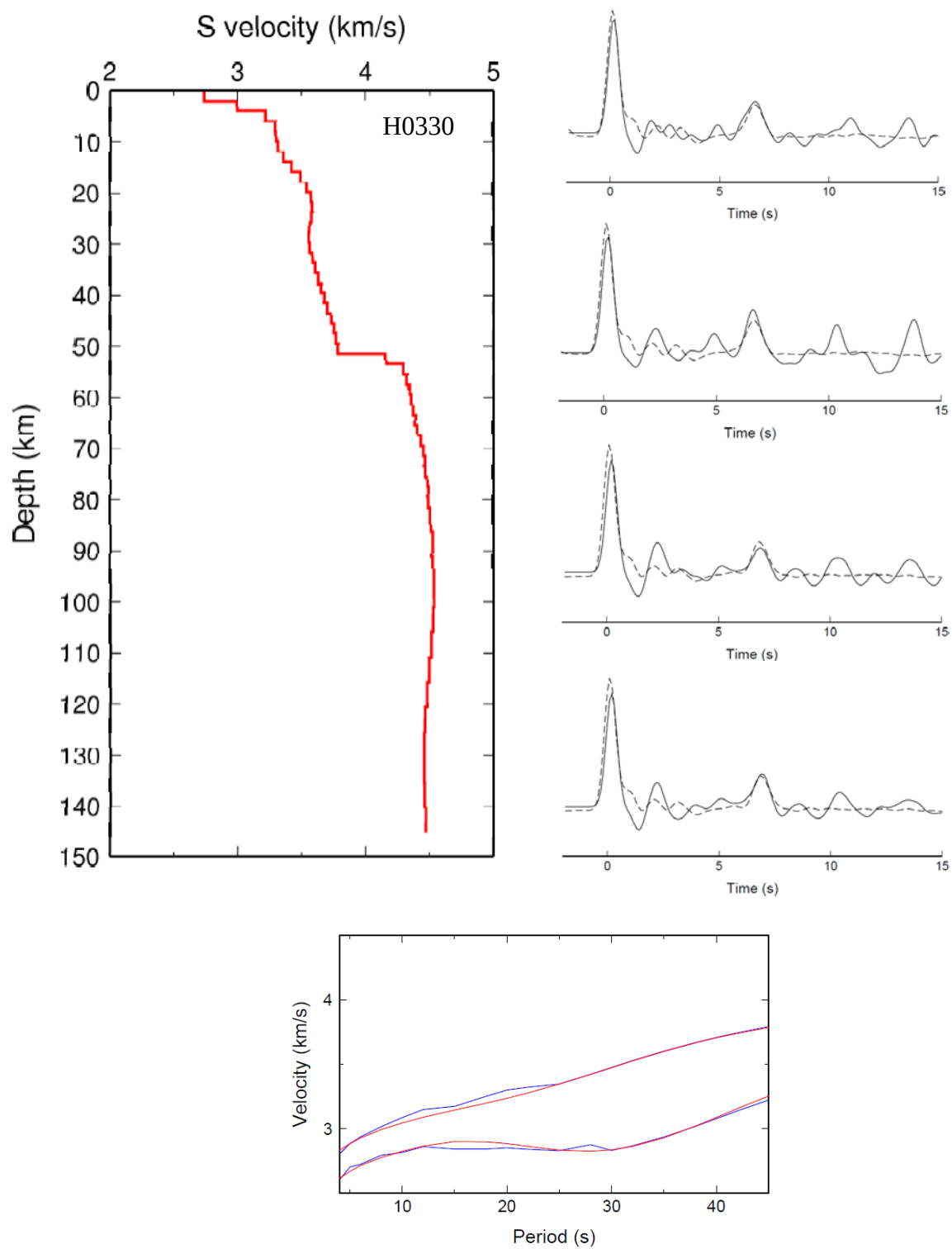
(a)

Fig. 3.4



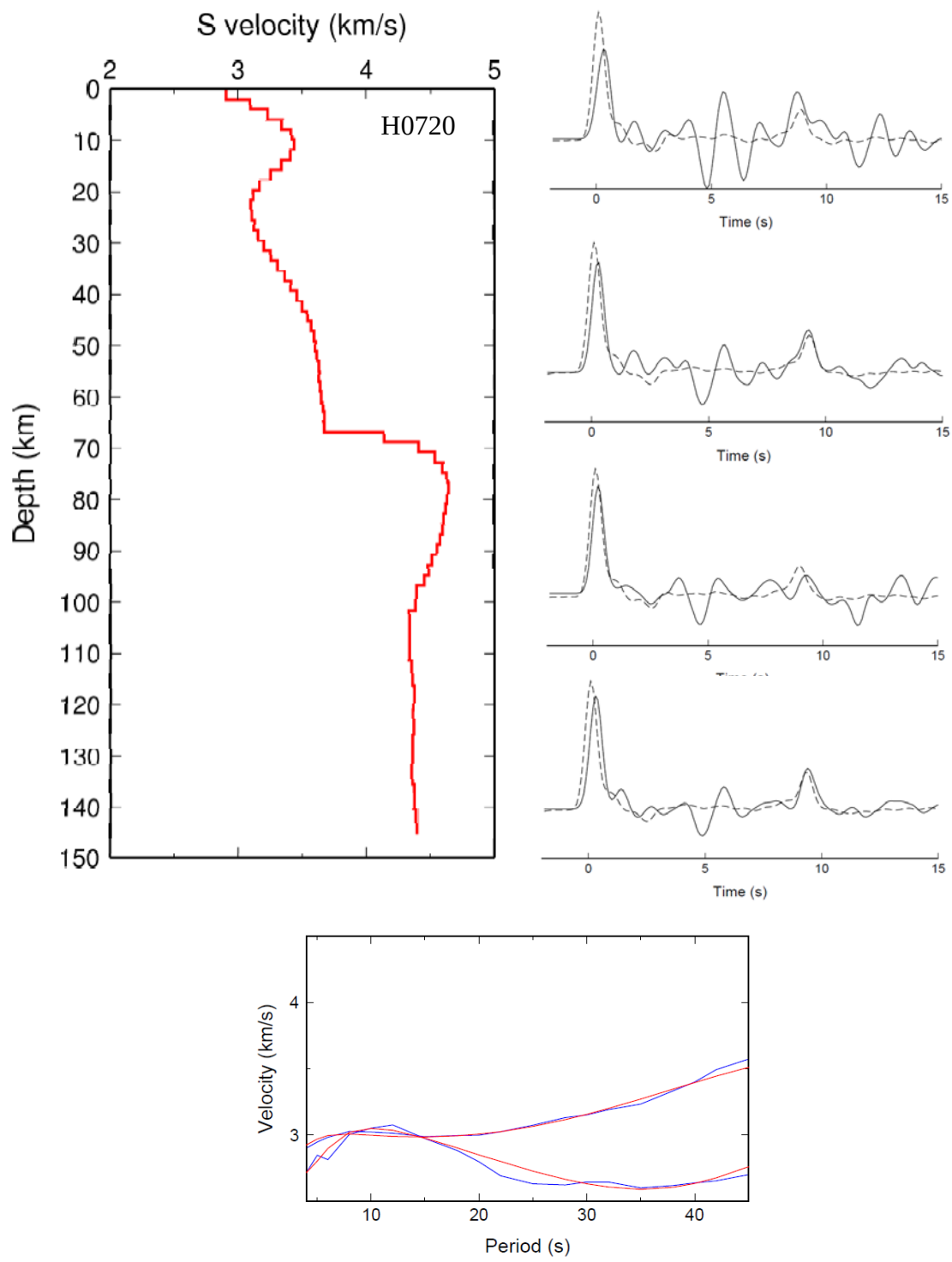
(b)

Fig.3.4. S velocity profile from joint inversion of receiver function and surface wave dispersion as projected to the red line in Fig. 3.1. (a) Color image of shear velocity structure. Topology of Tibetan Plateau is shown on the top of the figure. Crosses and circles indicate the Moho depth under each station from joint inversion. Crosses indicate stronger Moho discontinuity while circles mark the station with gradual Moho velocity transition. The thick white line is the approximated Moho location from the receiver function study by Nabelek et al. (2009) using the same Hi-CLIMB stations. The white dashed line indicates the location with weak Moho Ps conversion. (b) Assemble of 1D velocity profiles under different stations. One station is chosen from every three consecutive stations. Grey vertical line in the background shows the location of chosen stations.



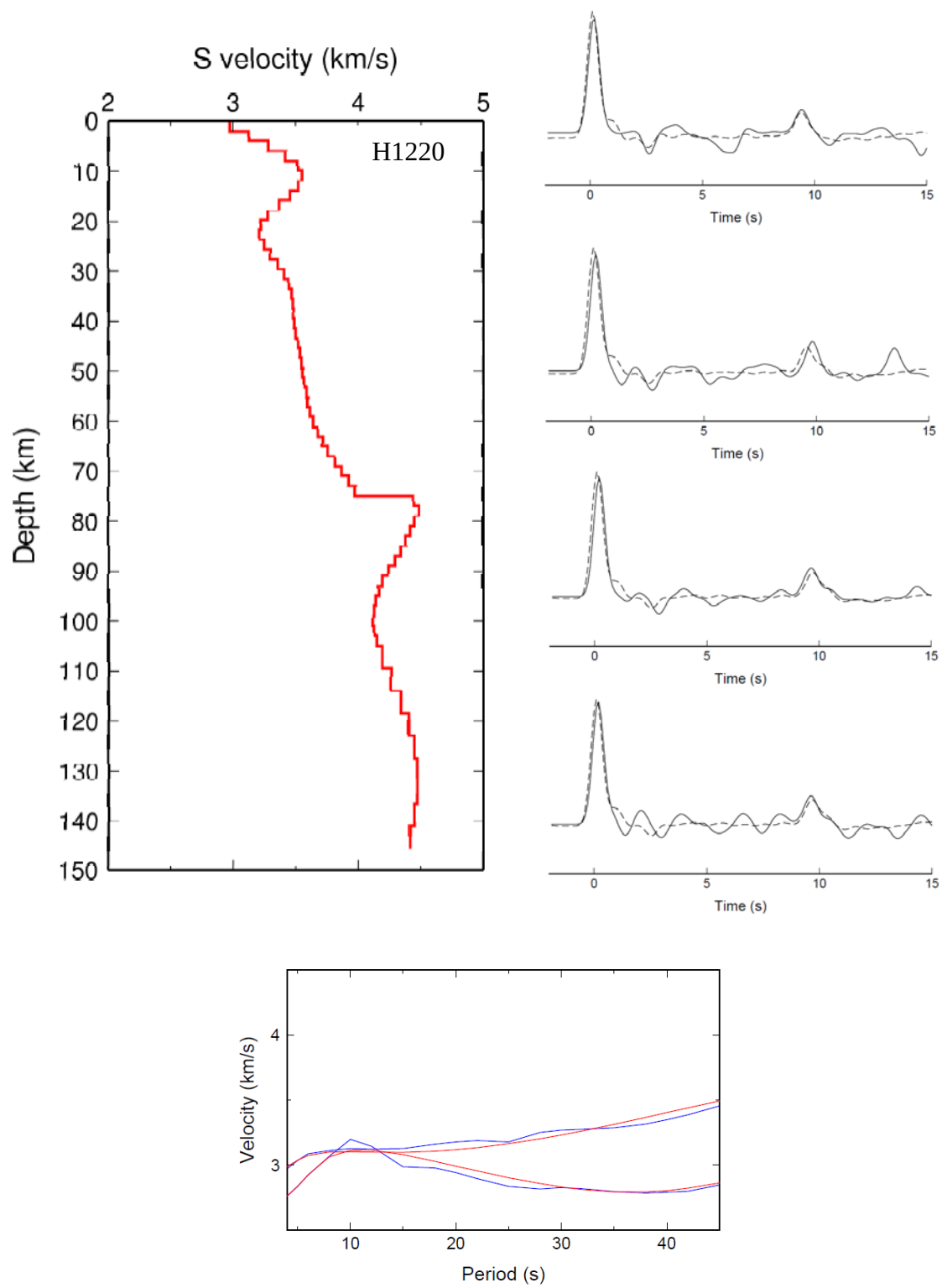
(a)

Fig. 3.5



(b)

Fig. 3.5



(c)

Fig. 3.5

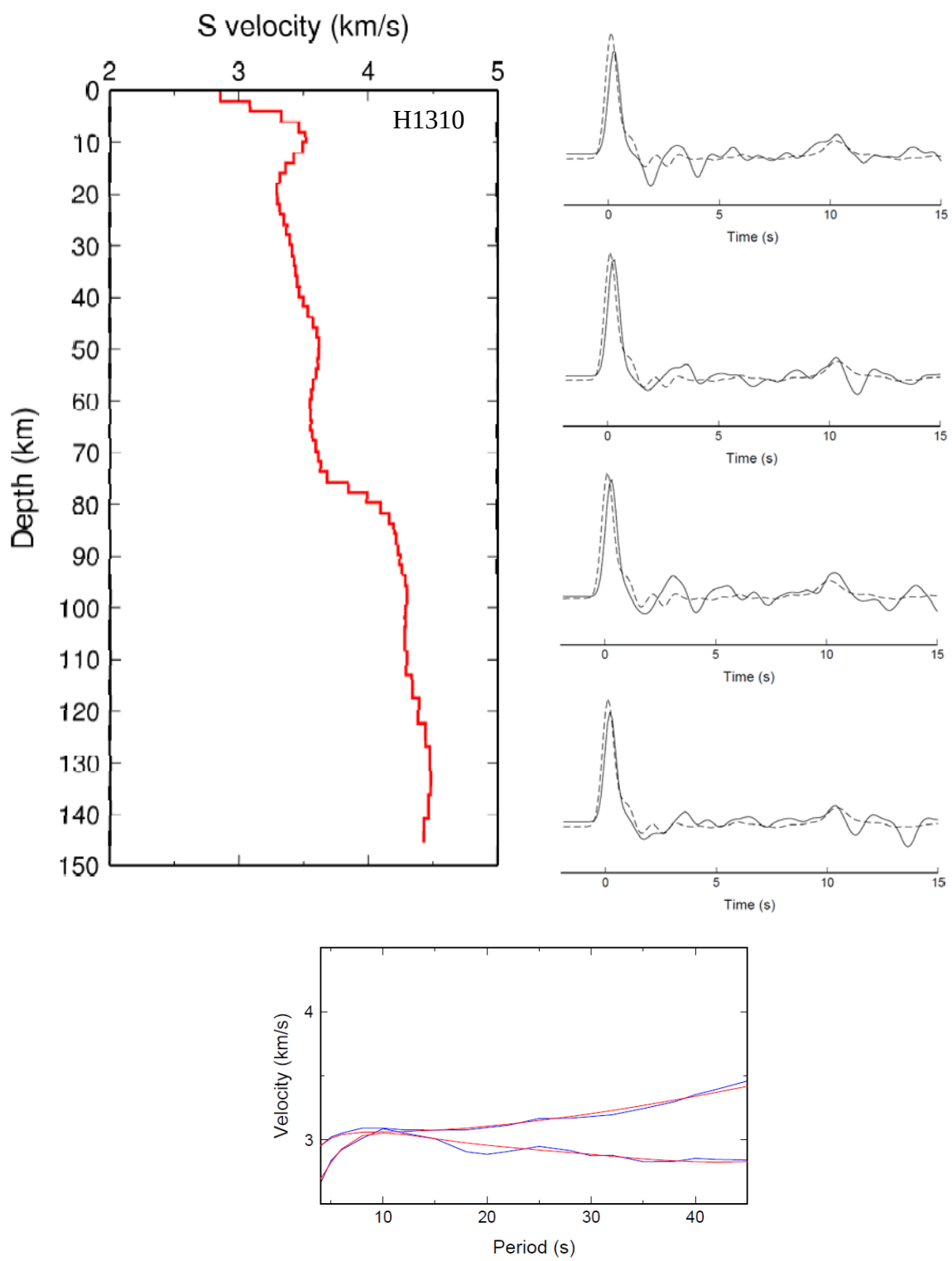
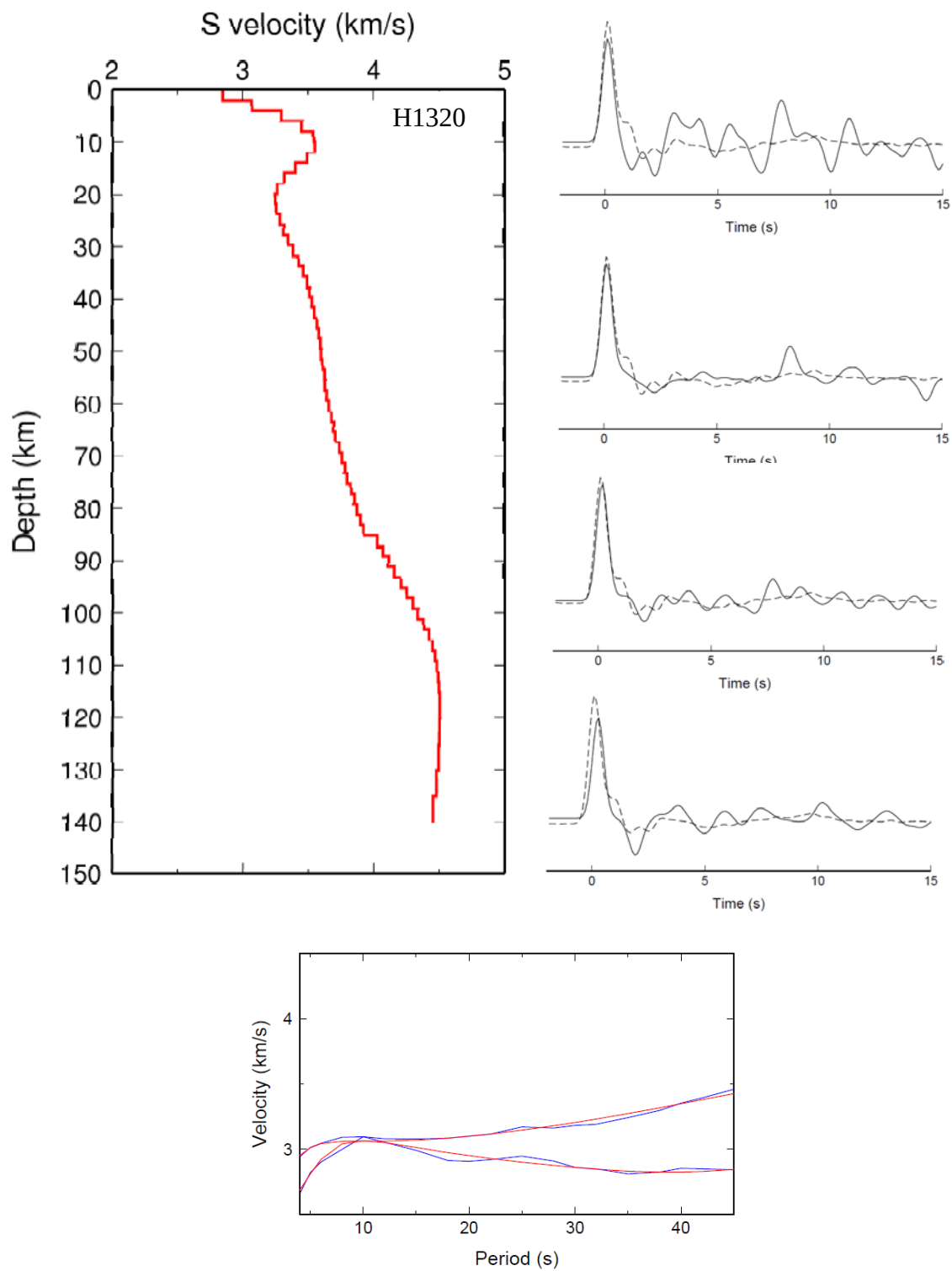
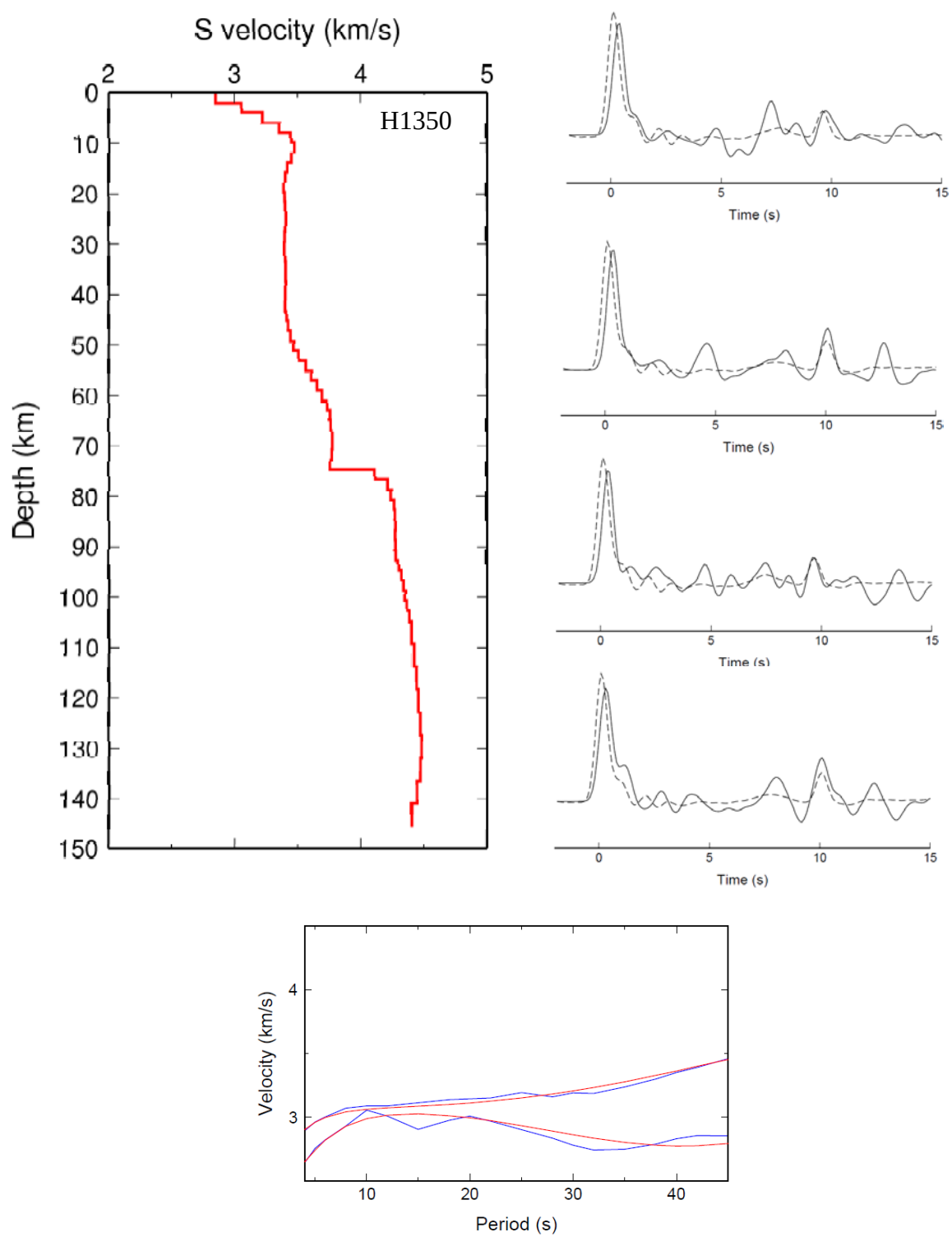


Fig. 3.5



(e)

Fig. 3.5



(f)

Fig. 3.5

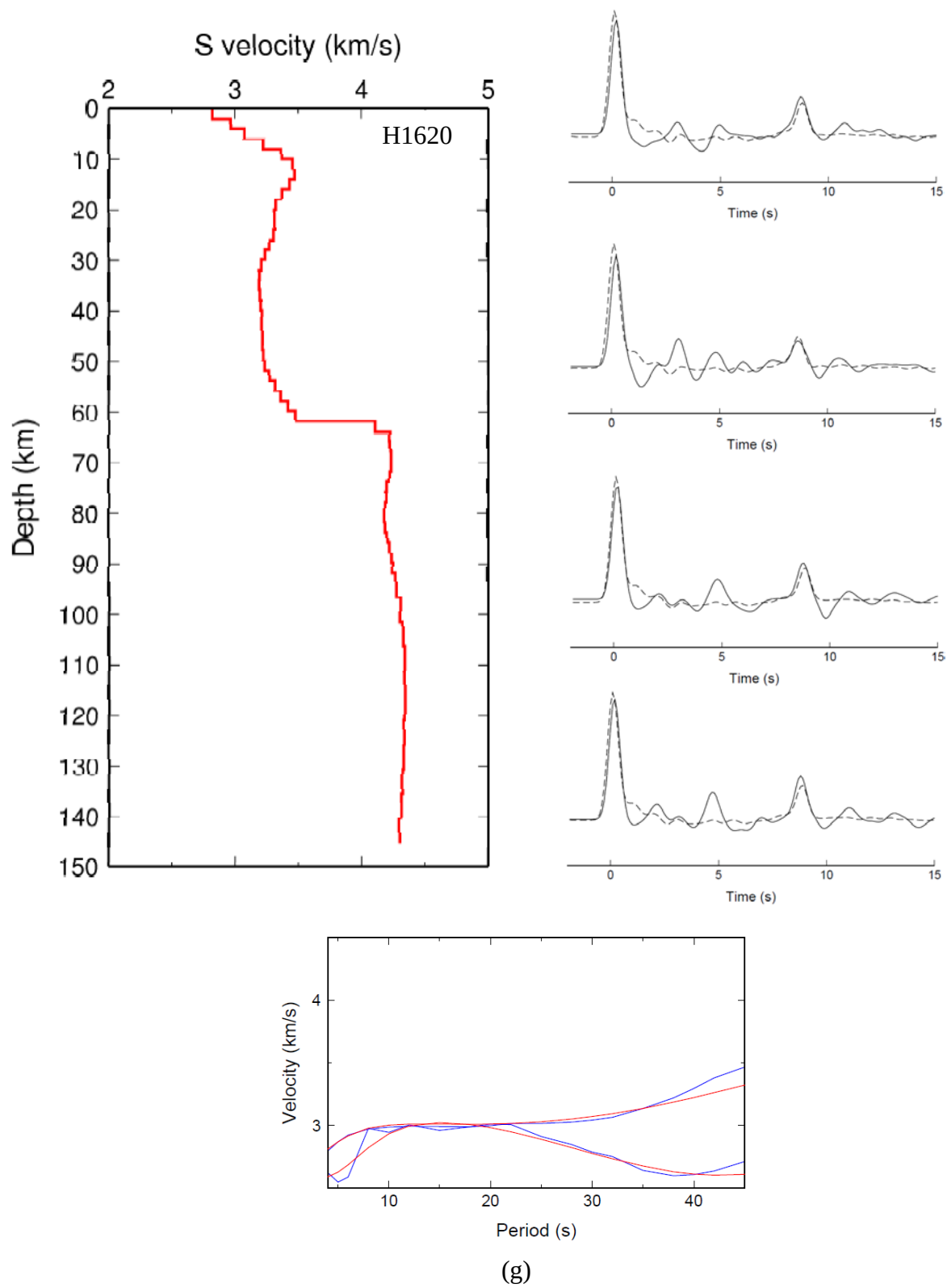


Fig. 3.5

Fig. 3.5. Surface wave dispersion and receiver function data and joint inversion results under Hi-CLIMB station H0330 (a), H0720 (b), H1220 (c), H1310 (d), H1320 (e), H1350 (f), H1620 (g), from south to north of the profile, respectively. The location of these stations is indicated in purple triangles in Fig. 3.1. Receiver function data are shown in black solid curve and predicted waveform are in dashed line. Dispersion data is plotted in blue color and predictions are in red.

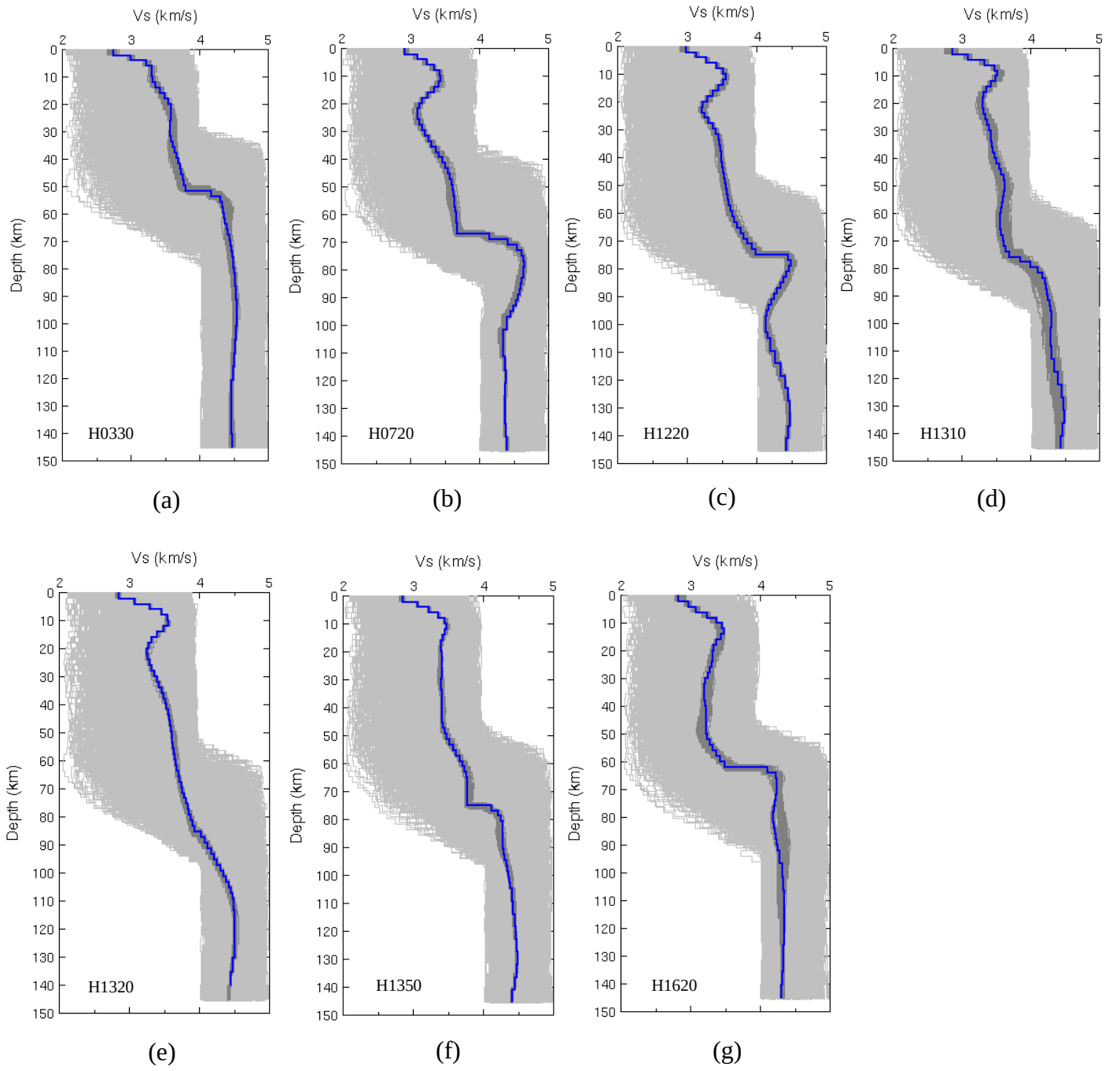


Fig. 3.6. Sampled models for each example station. Light greys in the background in each panel indicate a total of 80200 sampled models. Dark greys are 200 best fit models with smallest misfit value. Thick blue line is the best fit model for each station.

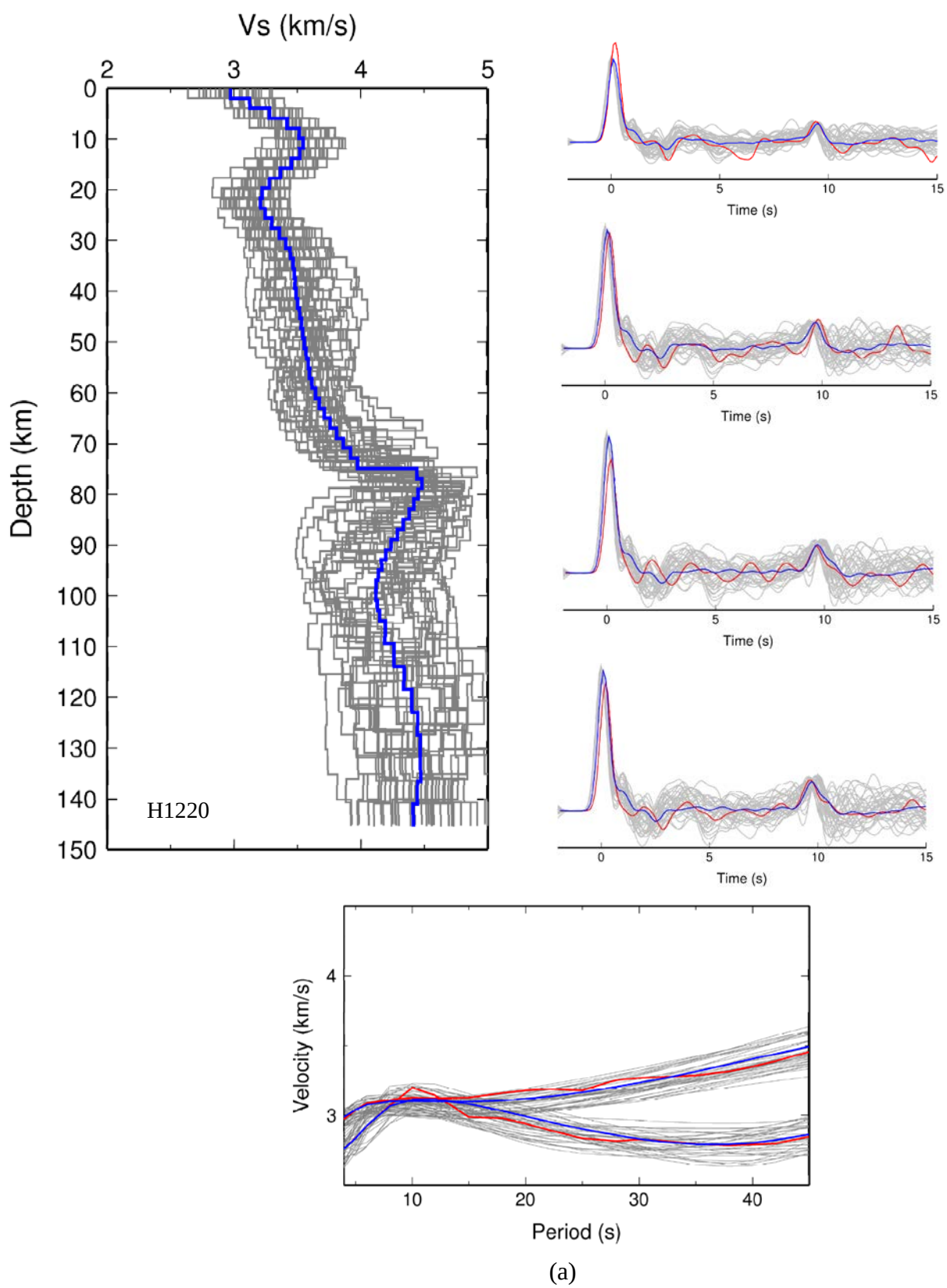
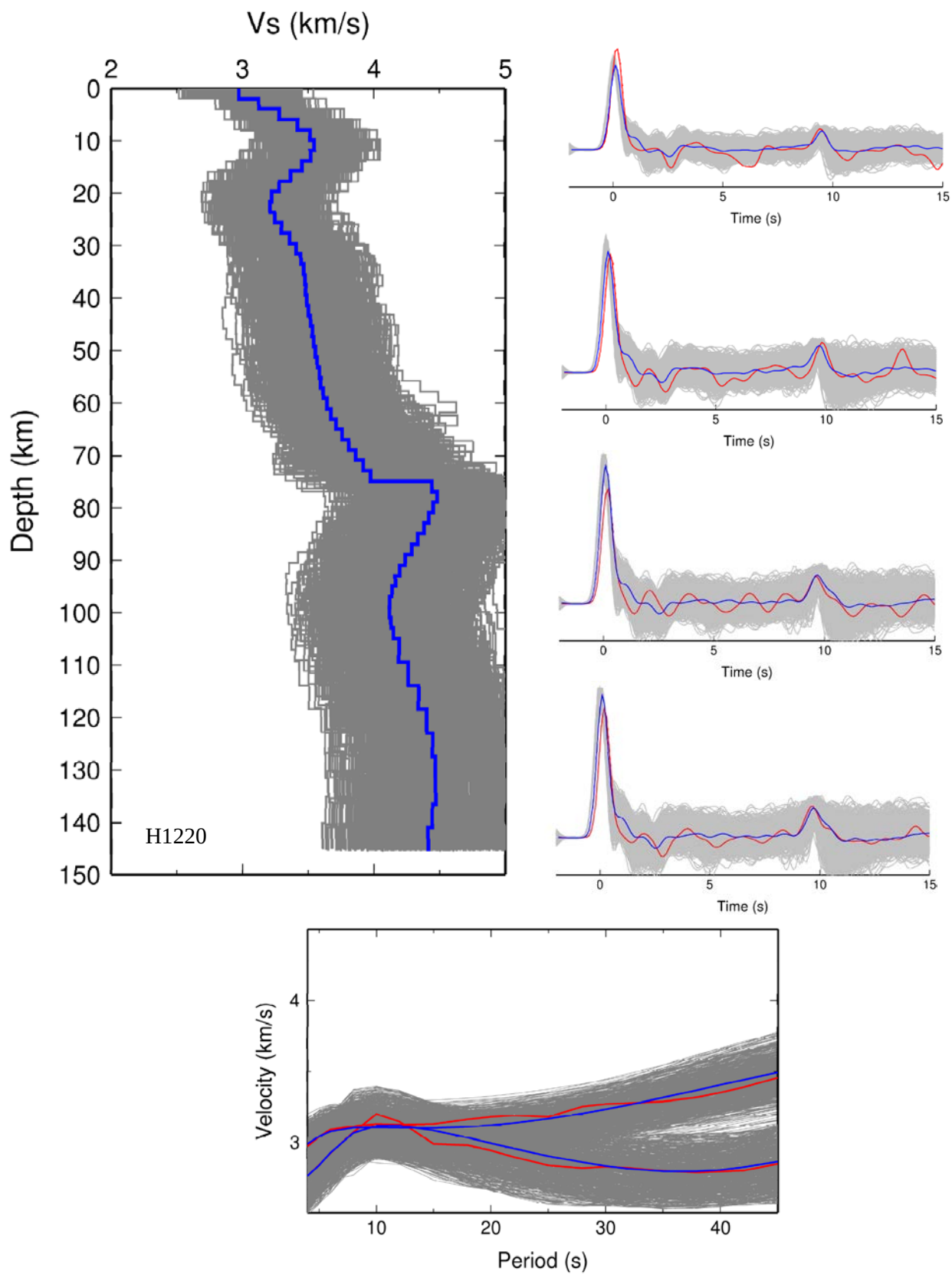


Fig. 3.7



(b)

Fig. 3.7

Fig. 3.7. Model sensitivity tests to the data. (a) Grey lines in the left panel are selected models from 80000 randomly generated models. The blue line is the best fit model from inversion. Grey lines in the right panels shows the predicted receiver functions and surface wave dispersions. Blue lines are the predictions from best fit model and red lines are the original data. The predictions of these models have Ps-P delay time difference less than 0.2 s with data; the difference of dispersions and maximum P and Ps amplitude with data is less than 6% and 10%, respectively. (b) Similar to (a) but with loosened selecting criterias (see text). Note a significant larger amount of models is selected and the predictions also cover a larger range.

CHAPTER 4

TEMPORAL CHANGES OF SURFACE WAVE VELOCITY ASSOCIATED WITH MAJOR SUMATRA EARTHQUAKES FROM AMBIENT NOISE CORRELATION

4.1 Introduction

Earthquake occurrence involves stress accumulation and sudden releasing of stored energy. Monitoring possible temporal changes of the medium before and after a major earthquake has been a long-sought goal for understanding earthquake genesis, recurrence, and prediction. Laboratory studies have long indicated stress-dependent elastic properties from microfracturing (e.g., ref. 1). However, because such a temporal change is likely small and the Earth structure is very heterogeneous, great care must be taken in this endeavor. Previous effort has relied on similar waveforms produced by repeated sources from earthquakes (e.g., refs. 2 and 3) or controlled sources (e.g., refs. 4 and 5). Because repeated sources are not regular in time or space, the application to monitoring has great limitations. Recent progress in passive imaging with random noise offers an alternative (e.g., reviewed in ref. 6).

Ambient noise has been widely used to retrieve empirical Green's function (EGF) of surface wave between stations (7, 8). The method involves cross correlation of continuous data for long enough time series. One major feature of ambient noise correlation is that the whole process is completely repeatable with different time segments, naturally producing EGFs with similar waveforms. Thus, the process makes it possible to continuously monitor the temporal evolution of the seismic wave velocity of the medium by tracking the time shifts of the EGFs.

The potential of this method has been shown by some recent studies. Sens-Schonfelder and Wegler (9) detected strong seasonal variation at Merapi Volcano, Indonesia from precipitation. Wegler and Sens-Schonfelder (10) detected a sudden drop in seismic velocity of 0.6% with the occurrence of Mw 6.6 Mi-Niigata, Japan earthquake from the autocorrelation of a station near the source. Brenguier et al. (11) reported precursory signals of less than 0.1% velocity variations before volcanic eruptions on La Reunion Island.

In this paper, we examine the 3 largest earthquakes in Sumatra, Indonesia in recent years (Fig. 4.1), occurring on December 26, 2004 (Mw=9.0); March 28, 2005 (Mw=8.6); and September 12, 2007 (Mw=8.5), respectively (magnitude from Global CMT solutions). The fault rupture length

ranges from 450 km (2007 event) to 1,200 km (2004 event) (refs. 12–14 and <http://earthquake.usgs.gov/regional/world/historical.php>). By comparing inter-station EGFs, we observed clear temporal changes in surface wave velocities after the earthquakes over a large area.

4.2 Results

4.2.1 Sumatra 2004 and 2005 Events

The available stations for 2004 and 2005 in the region are limited. We found 1 station pair, PSI in Indonesia and CHTO in Thailand, which has consistently high signal to noise ratio (SNR) in the EGFs for all of the months (Fig. 4.2). We examine time shifts (relative to reference EGF) of both Rayleigh wave (from cross correlation of vertical component) (Fig. 4.3A) and Love wave (from cross correlation of tangential component) (Fig. 4.3B) between PSI and CHTO. Detailed data processing is given below. The Love wave is very stable; throughout the whole time period, the time shifts are all within ± 0.5 s with the average and the standard deviation of -0.00 s and 0.17 s, respectively. For most time periods, the Rayleigh wave is also stable with the time shifts ranging from about -0.5 s to 0.5 s; the average and standard deviation are -0.09 s and 0.25 s, respectively, for the months that exclude the period from December, 2004 to April, 2005. However, we observe significant time shifts around the time of 2004 and 2005 events in the Rayleigh waves (Fig. 4.3A). The largest shifts are for the months right after the events. The shifts are 1.06 s and 1.44 s for the month right after the 2004 event and 2005 event, respectively. These values are 4.6 and 6.1 times the standard deviation away from the average of the other periods. The time shifts are clearly visible from the EGF waveforms (Fig. 4.2). In addition, there is a small DC shift between the data from May 2005 to 2008 and the data before the 2004 and 2005 events (Figs. 4.2 and 4.3A).

The fact that the time shifts of the Rayleigh wave during the whole time period that excludes the event months are very stable and are of the same amplitude as the time shifts of the Love wave suggests that the anomalous behavior right after the 2004 and 2005 events is unlikely to be random. Similarly, we found no evidence of any sudden change of ambient noise sources by inspecting the amplitudes and waveforms of the positive and negative delay parts of the inter-station correlations over calendar months (Fig. S1). To reinforce this conclusion, we show

another example from a station pair (HIA and MDJ in northeastern China) far away from Sumatra (Fig. 4.3C). The pair also has consistently excellent Rayleigh-wave EGFs for all of the months. The time shifts range from -0.5 s to 0.5 s for all of the months with the average and standard deviation of -0.02 s and 0.17 s, respectively. Overall, the time shifts show very similar scatter to the Love and Rayleigh waves of the CHTO-PSI pair outside the 2004 and 2005 event months, but no significant time shifts occurred in the 2004 and 2005 event months for the HIA-MDJ pair.

One problem may be clock error. A relative clock shift between the 2 stations will result in a time shift in the EGF from the ambient noise correlation. Clock errors can be examined using positive and negative delay in the EGF correlation (15). However, our EGFs are highly asymmetric. We thus compare time shifts of Rayleigh waves and Love waves. Because the horizontal components and the vertical component use the same clock for the same station, a clock error would affect the EGFs of both Rayleigh and Love waves. In addition, the time shifts appear frequency-dependent (see below), which cannot be caused by a clock error. Thus, clock errors can be ruled out as the cause for the large time shifts in the Rayleigh wave of CHTO-PSI during the time periods of the 2004 and 2005 event from the lack of similar shifts in the Love wave.

It is also noticeable that CHTO-PSI pair does not show significant time shift after September, 2007 event. This may be explained by the distance dependence of the media velocity change. Significant time shifts are observed in pairs closest to this event as described below. The 2007 event is further to the south thus the impact of media velocity change to CHTO-PSI pair is smaller. Moreover, the distance dependence may also account for the larger shift after the 2005 event, relative to that after 2004 event, even though the 2004 event has a larger magnitude.

We conclude that the large time shifts in Rayleigh wave after the 2004 and 2005 events are the result of temporal change of the velocity of the medium associated with these major earthquakes.

4.2.2 Sumatra 2007 Event

More stations are available in this region for the 2007 event with the addition of stations in Indonesia and Malaysia. To enhance SNR and to obtain more pairs, we compute the EGFs using 2 consecutive months of data with a moving step of 1 month. We calculate EGFs for all possible

station pairs and obtain 17 pairs with good SNR for all time segments between June, 2007 and June, 2008 (Fig. 4.1).

We observe clear time shifts after the earthquake for some pairs. The pairs related to the station MNAI closest to the epicenter show largest time shifts. There are 5 pairs related to MNAI for the September, 2007 event, i.e., KOM, PMBI, PSI, SBM, and UGM; the largest time shifts for these pairs are -0.26 s, 1.36 s, -0.43 s, -0.22 s, and -0.77 s, respectively. The pair closest to the epicenter of the 2007 event (MNAI-PMBI) shows the largest time shift (1.36 s), which occurred in the period of October–November, 2007 (Fig. 4.4). The largest time shifts for all of the other pairs occurred in the period of November–December, 2007 (the time shift for MNAI-PMBI in this period is 0.84 s).

The number of station pairs allows us to construct the pattern of time shifts for each time period by spatial averaging and to examine how the spatial pattern changes with time. We use a moving window of 10° by 10° in longitude and latitude and a step of 2° in longitude or latitude direction in the averaging. If there are 3 stations inside the window, we calculate the root mean squares (rms) of the time shifts for all of the pairs associated with the stations inside the window and find the averaged latitude and longitude of these stations. This rms value is then taken as the time shift at the averaged latitude and longitude for that time period. We perform the same spatial averaging for each time period and thus, construct a time evolution of the spatial pattern of the time shifts. Maps of the spatial pattern for a few months before and after the event are shown in Fig. 4.5, and Movie S1 shows the time evolution. It is important to stress that because we have carefully selected our station pairs so that each station pair selected has consistently good EGFs that allow us to measure the time shifts of each pair for all of the time periods, the spatial averaging has the same station-pair distribution and the same ray path coverage for all of the time periods. This data selection is important in eliminating any bias from non-uniform distribution of station pairs on the temporal evolution.

The rms values are small (less than 0.27 s) for the periods up to the period of July–August, 2007. However, strong anomalous signals appear after the earthquake. The strongest anomaly (up to 0.68 s) in the averages is near the epicenter, but the anomalies also appear 500 to 1,000 km away from the epicenter. The signals disappear further away from the source. The anomalies die away

~3 months after the earthquake and no significant anomalies are observed in any location after that. The evolution of the spatial pattern and the coincidence of both the location of the anomalies and the time when the anomalies emerge suggest strongly that the temporal changes are associated with the 2007 earthquake. In this interpretation, the earthquake caused changes in medium velocity over a large area near the source for a few months after the earthquake, resulting in the observed temporal changes in the Rayleigh wave travel times.

4.3 Conclusion and Discussion

We observe clear travel time shifts of Rayleigh wave in the months after the 2004, 2005, and 2007 major Sumatra earthquakes from ambient noise correlation. The absence of anomaly in Love wave suggests the Rayleigh wave anomaly is not from clock error. High SNR and similar waveforms of the EGFs for different periods ensure the accuracy of the measurements of the time shifts. The stability of the background measurements suggests the significance of the anomalous time shifts. The locations and emerging times of the anomalies coincide with the 3 major events. Therefore, we conclude that the travel time anomalies of Rayleigh wave are from the temporal change of surface wave velocity from a change of medium velocity after the major earthquakes.

The discrepancy between the Rayleigh and Love waves is intriguing. Because Rayleigh wave is sensitive to deeper structure than Love wave at the same period, 1 possibility is that the temporal change occurs at depths below the Love wave sensitivity. We examine this issue by determining whether the time shift is frequency dependent. To increase the SNR for the frequency analysis, we stacked correlation of CHTO-PSI in the 2 months right after 2004 and 2005 events. We filter the stacked correlation with 7 different period bands (5–10 s, 7–12 s, 9–14 s, 11–16 s, 13–18 s, 15–20 s, and 17–22 s, respectively). We measure the relative time shift with respect to the reference EGF for each period band. Because the surface wave is dispersive, we carefully select different time windows for different periods to include all of the surface wave energy (Fig. 4.6A). Our measurements show that the time shift increases and then decreases with period with the maximum shifts at 11–16 s (Fig. 4.6B). Measurements using 1 month of data after the 2004 and 2005 events, respectively, confirm the trend, although the SNR is not as good at marginal periods. The depth region of maximum sensitivity of fundamental Rayleigh wave varies from 10 to 30

km for period 10–20 s, which is in agreement with the depth range of maximum slips of 2 events (14). However, fundamental Love wave for this period range is sensitive to shallower structure (maximum sensitivity less than approximately 10 km). This depth dependence of the media velocity change can account, at least partly, for the discrepancy between the Rayleigh and Love waves, particularly if the media change is concentrated within small depth range of maximum rupture, for example. Alternatively, the medium velocity change may be anisotropic, resulting in different behaviors of Rayleigh and Love waves. More data are needed to sort out precisely the nature and the location of the media change.

Our observed temporal changes are from station pairs separated by several hundreds to 1,000 km and from 10–20 s surface waves sampling the depth of upper-mid crust (10–20 km), indicating a large-scale disturbance of the medium beyond the rupture areas. The 3 events are thrust events with small dip angles dipping toward the Sumatra coast. The imaging of the rupture areas (14) and the aftershock activities indicate that the west most stations are near the edge of the ruptures (PSI to 2004 and 2005 events and MNAI to 2007 event) and all of the station pairs are away from the rupture areas. The observed velocity change of the medium seems complex. Both positive and negative time shifts are observed after the 2007 event when several station pairs are available. There is no obvious correlation between the length of the path and the amplitude of time shift. Thus, the medium velocity can increase or decrease. The fluctuations are greater near the source to produce larger rms values as observed for the 2007 event.

Mechanisms that have been proposed for temporal changes of seismic velocity include (i) stress-induced changes of fault zone properties at seismogenic depth (3, 5, 10, 16), (ii) damages from fault zone rupture and subsequent healing (4, 17), (iii) damages of shallow crust from strong-ground shaking (18, 19), and (iv) rapid changes in ground water near the surface (9) or fluid activities in shallow crust (20). The depth sensitivity of the observed Rayleigh wave temporal changes suggests that surface or shallow crust effects (mechanisms 3 or 4) can be ruled out. As our station pairs are far away from the earthquake rupture areas, we also argue that the temporal changes are not directly related to the rupture damages or other properties within the fault zones. The most plausible explanation is stress changes associated with the major earthquakes and subsequent relaxation. The stress changes can be observed not only in the immediate vicinity of the rupture area but also several hundreds of km away from the source region. Interestingly, the

“DC” residual time shifts (Figs. 4.2 and 4.3A) after the 2004 and 2005 events, relative to those before the events (with an average of -0.46 s from April–October, 2004), may be part of the long-term stress relaxation (16). On the annual base from May through April, the average time shift has generally decreased from 0.00 and 0.10 s in 2005–2007 (with a data gap from January–August, 2006) to -0.06 s, -0.21 s in 2007–2009 (with measurements extended to April, 2009). A longer-term monitoring is required to confirm the trend.

There appears a significant excursion before the 2004 event in the CHTO-PSI pair (Fig. 4.3A). To examine the stability of the time-shift measurements, we construct EGFs for a moving window of 1 month and a moving step of 10 days for the period of April, 2004 to February, 2005 (no data are available before April, 2004) and obtain the time shifts relative to the reference EGF as before (Fig. 4.3D). The time shifts from April, 2004 to October, 2004 are very stable: the average is -0.443 s and the standard deviation is 0.123 s. However, the time shift (0.09 s) for the period November 6 to December 6 increases from the average of the previous months by 4.3 times the standard deviation of the previous months. The curve decreases a little and then increases sharply after the event (December 26, 2004).

The above statistics are affected by the limited duration of observations before the 2004 event. Of particular concern are seasonal variations in time shifts, such as those found by ref. 10, where velocity perturbations were found to correlate with precipitation, and thus can be explained by seasonal variation of underground water. However, at period 10–20 s, our Rayleigh waves are not sensitive to superficial conditions. If we compare variations of time shifts from year to year, there is no seasonal signal (Fig. S2). There are, however, short-term excursions from all of the months at which we have looked (April, 2004 through April, 2009, excluding the months where there are no data and those of Sumatra earthquakes January–April, 2005), among which the November 2004 anomaly is the largest. If we compare each month with the previous 6 months where there are measurements, we can find 3 strongest excursions (November, 2004 departing by 4.2σ , January, 2007 by 3.7σ , and March, 2008 by 3.1σ). In November and early December, 2006, there was only 1 magnitude >7 event (November 11, 2004, Timor Region, Mw 7.5, Lat. -7.87°, Long. 125.12°). The event is too far away from CHTO-PSI path to cause observable signals. We conclude that the anomaly before the 2004 event is unusual, but we cannot rule out that it is a random excursion.

Our observations of large-scale disturbances in medium velocity before and after a major earthquake need to be confirmed with a denser network around the source region. Recent large earthquakes with dense instrumentation nearby (such as 2008 Wenchuan, China; 2002 Denali, Alaska; and 1999 Chi Chi, Taiwan) provide a fertile ground for testing. The large-scale disturbance observed in this study suggests the stress-induced change in elastic properties is not just concentrated in the fault-zone area, making it easier to observe. If a precursor of media velocity change is observable over a large area, it has direct application for earthquake prediction.

4.4 Data Processing

We requested 54 months of long period continuous data (January, 2004 to June, 2008) for stations in the Southeast Asia region (Fig. 4.1) from Incorporated Research Institutions for Seismology Data Management Center (IRIS DMC). The stations are operated by GSN, Malaysia, Japan, and China. We retrieve inter-station EGFs using a computer program (with some modification) from Michael Ritzwoller's group, described in details by Bensen et al. (21). We filter the raw data between 4 to 50 s. For each of the Sumatra events, we exclude the data of the event date to eliminate the strong earthquake signals in the correlation. The EGFs obtained from the ambient noise correlation for almost all station pairs are not symmetric between the positive time delay and the negative time delay. Examples of ambient noise correlations are shown in Fig. S1 for the CHTO-PSI pair. The asymmetry indicates the noise sources are mostly from the south (Indian Ocean). Note the relative amplitude between the negative delay (containing Rayleigh wave EGF) and positive delay (mostly noise) is quite stable, suggesting no systematic or sudden change of the ambient noise source and that Rayleigh wave EGFs can be consistently retrieved. We use the symmetric part of the EGF by summing both the positive delay and negative delay parts, which enhances the SNR and stability of the EGF. We then filter the EGFs with a band pass filter between 10 and 20 s, where SNR is highest for most station pairs.

To examine temporal change, we obtain EGFs from short time segments (e.g., 1 month) of continuous data. A reference EGF is constructed using all of the time periods. Only those pairs with consistent high SNR for each time segment throughout the whole time period were selected for analysis. We measure the time shift of the EGF from each time segment relative to the reference EGF using waveformcross correlation. As an example, Fig. 4.2 shows Rayleigh wave

EGFs between stations PSI (in Indonesia) and CHTO (in Thailand). The waveforms are highly similar, making it possible to measure accurately relative time shifts. We typically use a large time window (~ 60 s) that includes the large portion of the surface wave packet in measuring the time shift with cross-correlation method. The sampling interval of the retrieved EGFs is 1 s (same as the raw data). We interpolate the EGFs to sampling interval of 0.01 s before using waveform cross-correlation to measure the relative time shifts.

One major source that could contaminate the noise-correlated EGFs is the intensive aftershock signals, even though normalizations in time and frequency domains have been applied to remove any energetic signals in the EGF construction (21). To test the effect of aftershocks, we compare EGFs using 1 month of data after the 2004 event and 1 month of data after the 2005 events with and without major aftershocks. Both months show significant positive time shifts as discussed above. We mute the records 1 min before expected arrivals and 1 h after the expected arrivals for all of the events with Mw 5.5 and above in our study region (10° S to 20° N and 90° E to 110° E). Because earthquakes with Mw 5.5 are quite common in this region and there is at least 1 event happening in almost every month, we determine that Mw 5.5 is a reasonable cut off to study the behavior of major aftershocks. We windowed out a total number of 41 and 24 aftershocks found in the time period from December 27, 2004 through January 25, 2005 and from March 29, 2005 through April 27, 2005, respectively. To avoid artificial spikes, the data windowing is done after filtering of the raw data and before the temporal normalization (21) and a cosine taper (width of 120 s) is used in the data windowing. About 95% and 97% of data are kept for noise correlation after windowing for the month after the 2004 event and the month after the 2005 event, respectively. The waveforms of EGFs before and after aftershock windowing, the waveforms are almost exactly identical with only a slight change in amplitudes (Fig. S3), indicating that the Green's functions we retrieved are stable and the observed large time shifts after the 2004 and 2005 events are not likely from aftershocks.

Rayleigh waves can be retrieved from either correlation of vertical component (Z-Z) or correlation of radial component (R-R). The latter, however, is subject to errors in instrument orientation (azimuth). There is a long data break between December, 2005 and September, 2006 for PSI station and CHTO changes its orientation after September, 2006. Examination of Z-Z, R-R, and T-T noise correlations for the CHTO-PSI pair suggests that, while Z-Z and T-T

correlations are generally consistent throughout the time period, the waveforms of R-R correlation are not consistent with a marked change after the data break (after September, 2006) (Fig. S4). Love wave energy leaked to R-R correlation to become the dominant energy after the data break, suggesting significant error in instrument orientation. Because Rayleigh wave arrives later than Love wave, it may be contaminated by the coda of Love wave. Therefore, we used Z-Z correlation, rather than R-R correlation, for Rayleigh waves in this study. Nevertheless, using Rayleigh waves from R-R correlation from April, 2004 to December, 2005, where the waveforms are generally consistent, we found significant time shifts (0.87 s and 1.76 s, respectively, relative to the average before October, 2004) after the 2004 and 2005 events that are generally consistent with the results from Z-Z correlation. On the other hand, Love waves from T-T correlation in the CHTO-PSI pair are more consistent, which travels faster than Rayleigh wave; although there is also an apparent change in the coda of Love waves (Rayleigh wave energy) after the data break (Fig. S4B). The consistency of Love waveforms make it possible to measure accurately the time shifts as presented above.

4.5 References

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4.6 Figures

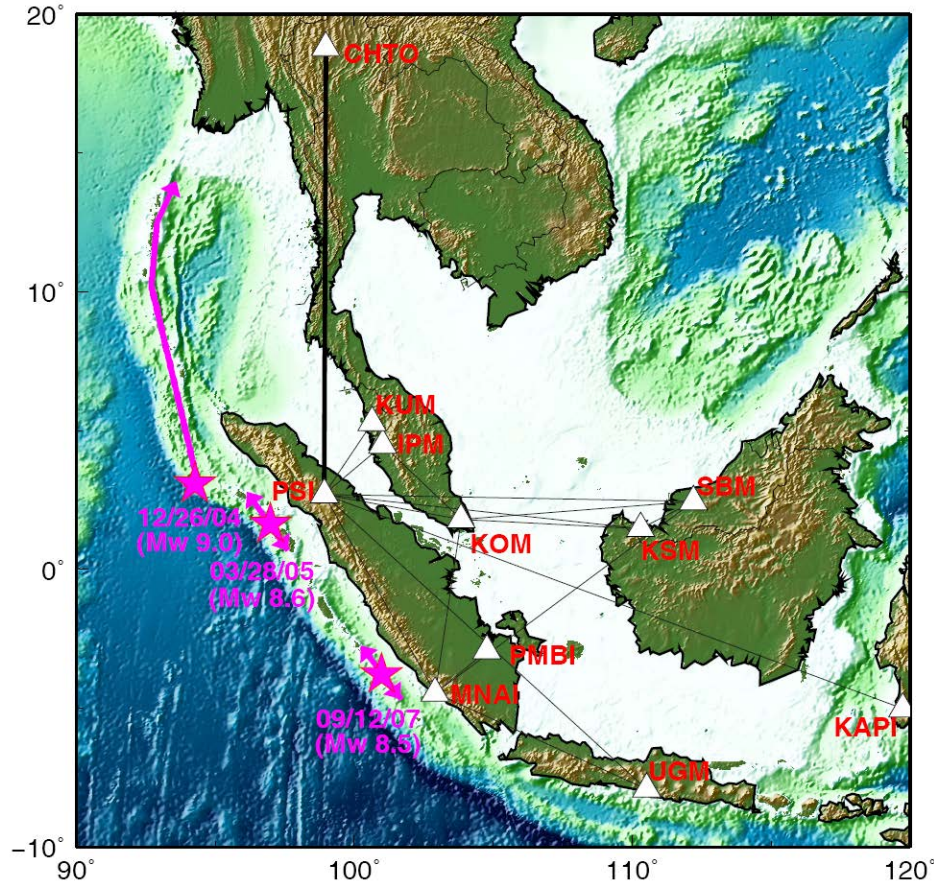


Fig. 4.1. Map showing the location of the 3 major Sumatra earthquakes (stars) and seismic stations used in this study (triangles). Lines indicate the station pairs that we have time shift measurements. The black thick line is the pair CHTO-PSI for the 2004 and 2005 events, and the thin lines are the pairs used for the 2007 event. Arrows represent the rupture direction of the earthquakes. Rupture areas are simplified from previous studies (refs. 12–14 and <http://earthquake.usgs.gov/regional/world/historical.php>).

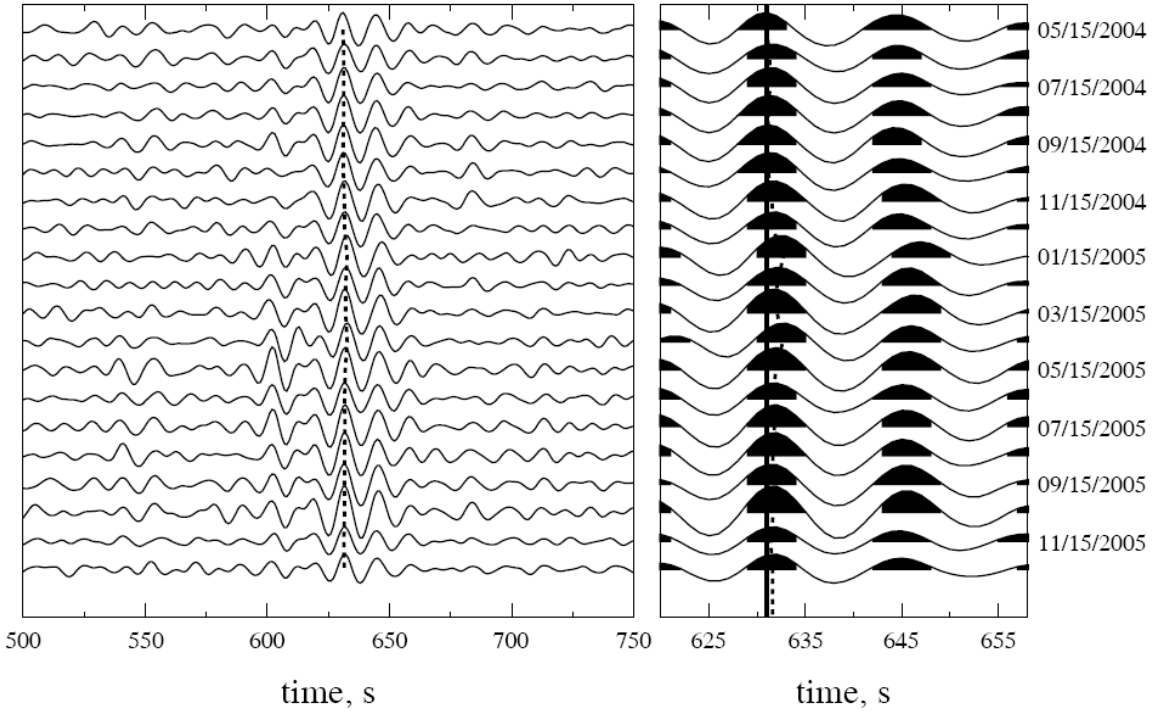


Fig. 4.2. Rayleigh waves of the station pair CHTO-PSI from ambient noise correlation for the period from May, 2004 to December, 2005. Each trace corresponds to 1 month of stack. All of the traces have high SNRs, and the waveforms from different months are highly similar. The enlarged view in the right shows clear time shifts in the months of December, 2004 to June, 2005. The difference between the reference line (solid) and the dashed line indicates the relative shift between traces.

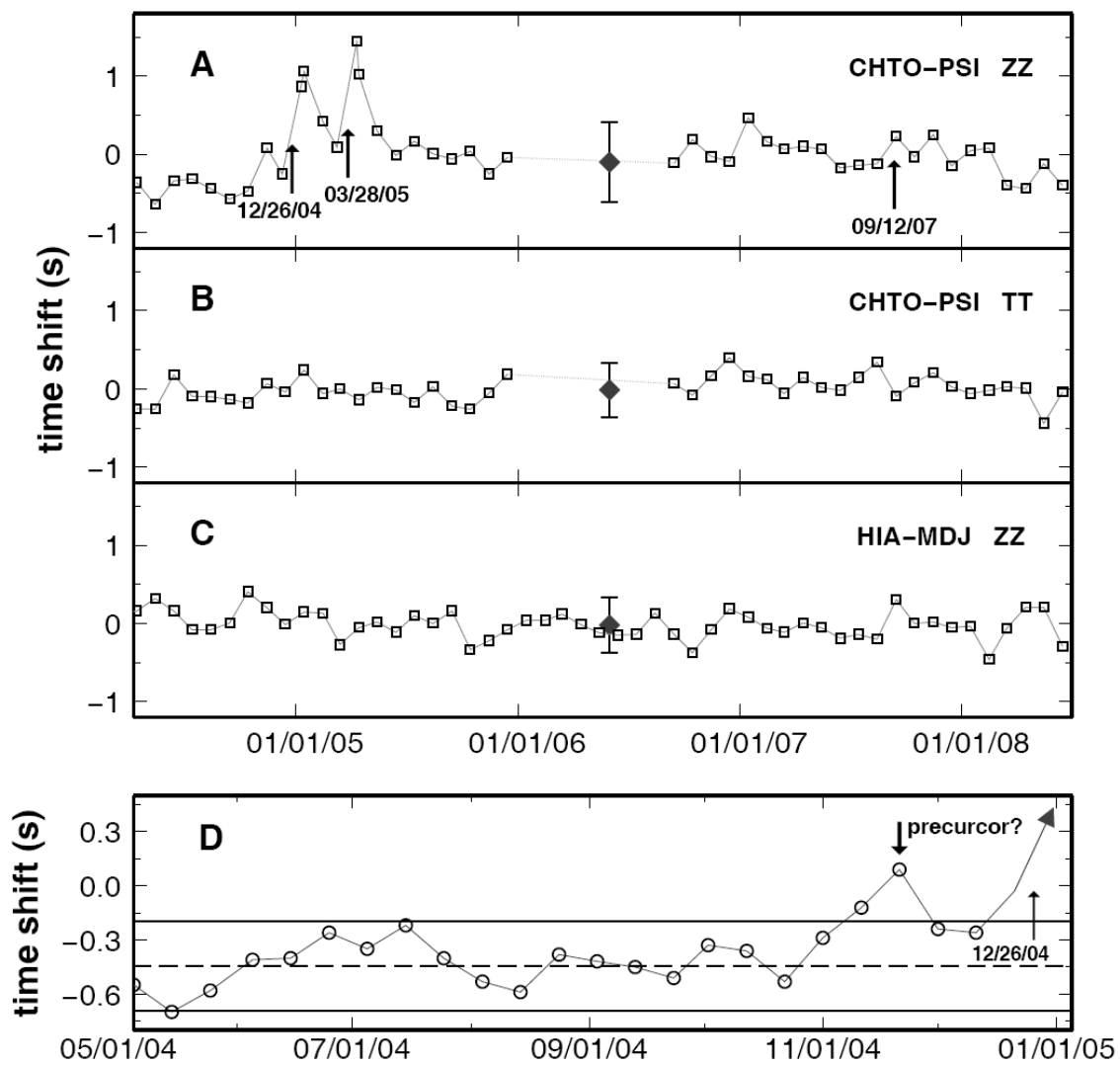


Fig. 4.3

Fig. 4.3. Time shifts of various station pairs, relative to reference stack of all time periods. (A–C) Rayleigh wave for CHTO-PSI from vertical (Z) component correlation (A), Love wave for CHTO-PSI from tangential (T) component correlation (B), and a stable reference pair HIA-MDJ outside Sumatra region in northeast China (C). Black open squares correspond to time shift in different calendar months. Every data point is put in the middle of that month. In A, 2 additional points were calculated using 1 month of data before and after the 2004 and 2005 events that exclude the event dates. Gray dashed lines in CHTO-PSI ZZ and TT plots indicate the time period from January, 2006 to August, 2006 when no data are available. Solid diamond with error bar indicates the average time shift ± 2 standard deviations for all of the time periods except 2 months after the 3 Sumatra events. Arrows mark the dates of the 2004 and 2005 events. (D) Enlarged view of time shifts of CHTO-PSI Rayleigh wave before December, 2004 event, showing a weaker but significant precursory signal. Dashed line indicates the average time shift in the period before 2004 event. Two solid lines mark the 2 standard deviations from the average. The time shifts (open circle) are associated with a Rayleigh wave that is calculated using every 30 days before the 2004 events (rather than using the calendar months). The time shift is plotted at the middle of each time period. The trend curve at the end uses measurements from A, which contain a mixture of data before and after the events.

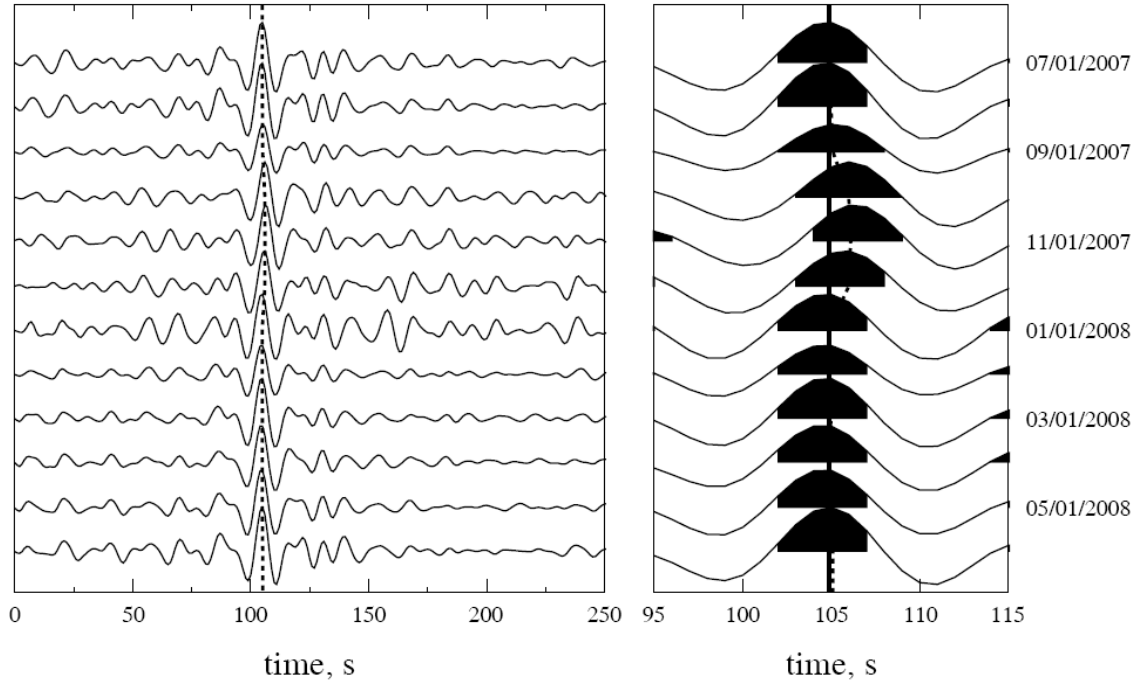


Fig. 4.4. Rayleigh waves of MNAI-PMBI pair from ambient noise correlation for the period from June, 2007 to June, 2008. Each trace corresponds to 2 months of stack. The traces are aligned in the middle of the 2 months; e.g., 07/01/2007 indicates the stack from June 1, 2007 to July 31, 2007. Clear time shifts are observable for the periods of September–October, 2007; October–November, 2007; and November–December, 2007.

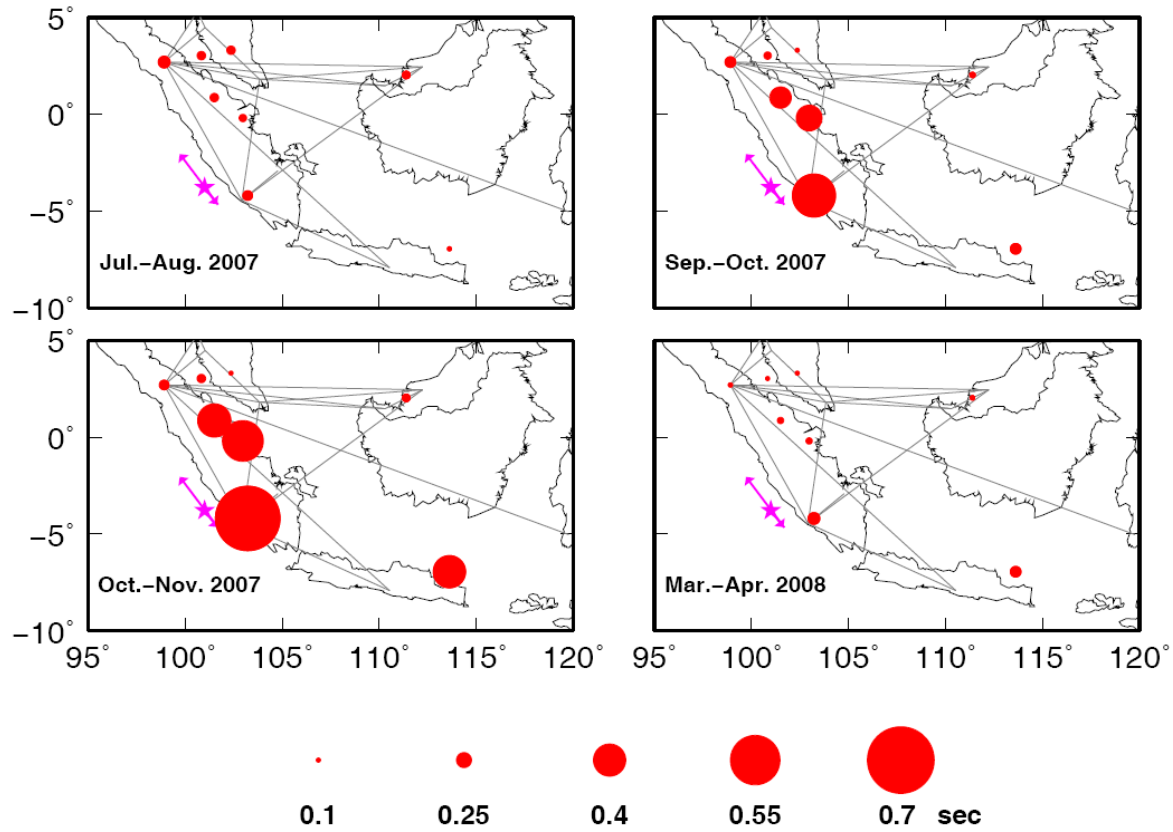


Fig. 4.5. Snapshots of time-shift maps before (*Upper Left*), during (*Upper Right*) and after (*Lower, Left and Right*) the 2007 event. The time period is indicated in each panel. For each time period, a 10° by 10° window with 2° of overlapping in both longitude and latitude is used for spatial averaging. The rms time shift is calculated in each window only if there are at least 3 pairs related to the stations inside the window. The red dot represents the rms time shift plotted at the average location of the stations inside each window. The size represents the relative amplitude of the rms time shift.

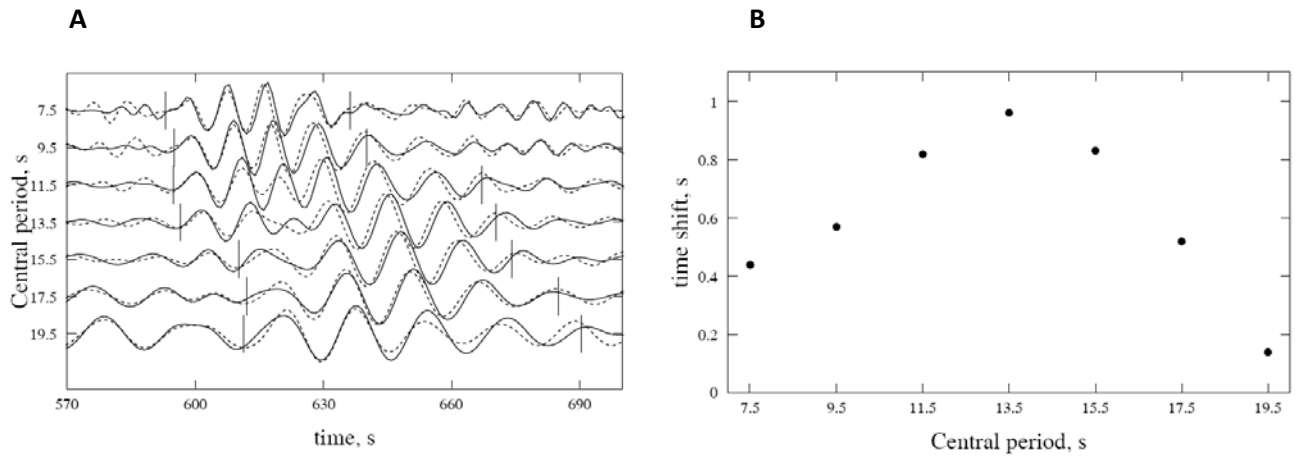
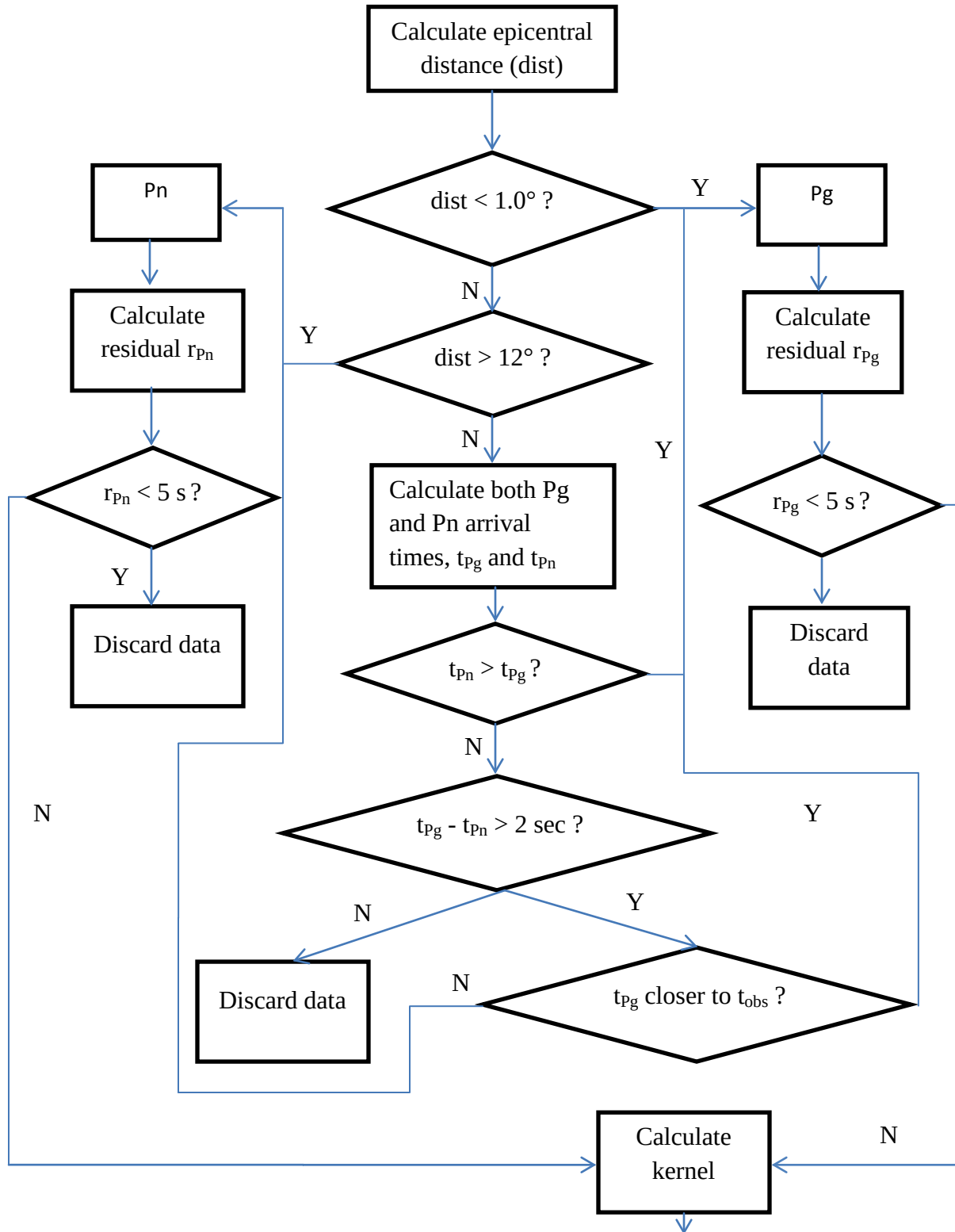


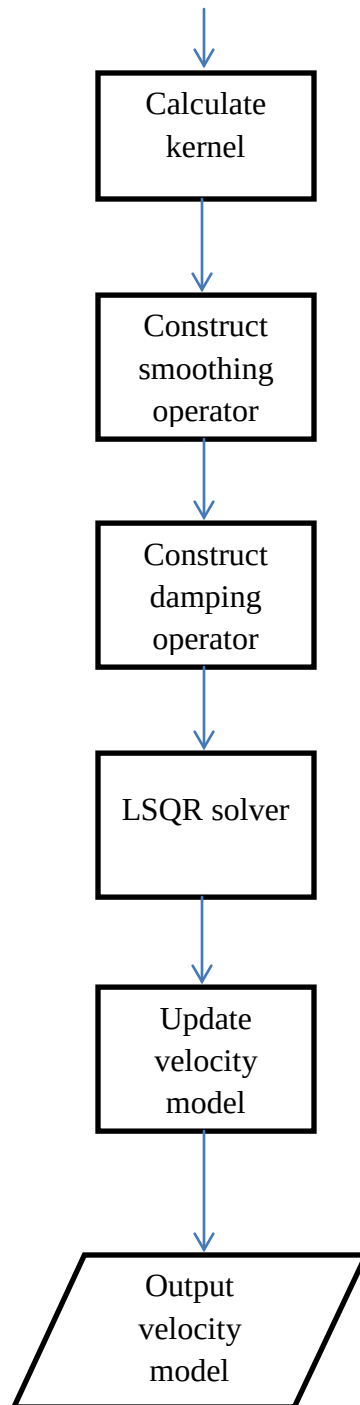
Fig. 4.6. Waveforms in different period bands (A) and corresponding time shift measurements (B) for CHTO-PSI pair. Two months of Green functions after 2004 and 2005 events are stacked and bandpass filtered to 7 different periods at 5-s intervals. Solid traces indicate the 2-month stack, while dashed traces are from the reference Green function. Each trace is normalized to its largest amplitude. The largest shift is observed in intermediate period bands. Solid vertical ticks in each trace indicate the time window we used to measure the time shifts.

APPENDIX A: SUPPLEMENTARY TO CHAPTER 1

A.1 Flow chart of 3D joint inversion scheme

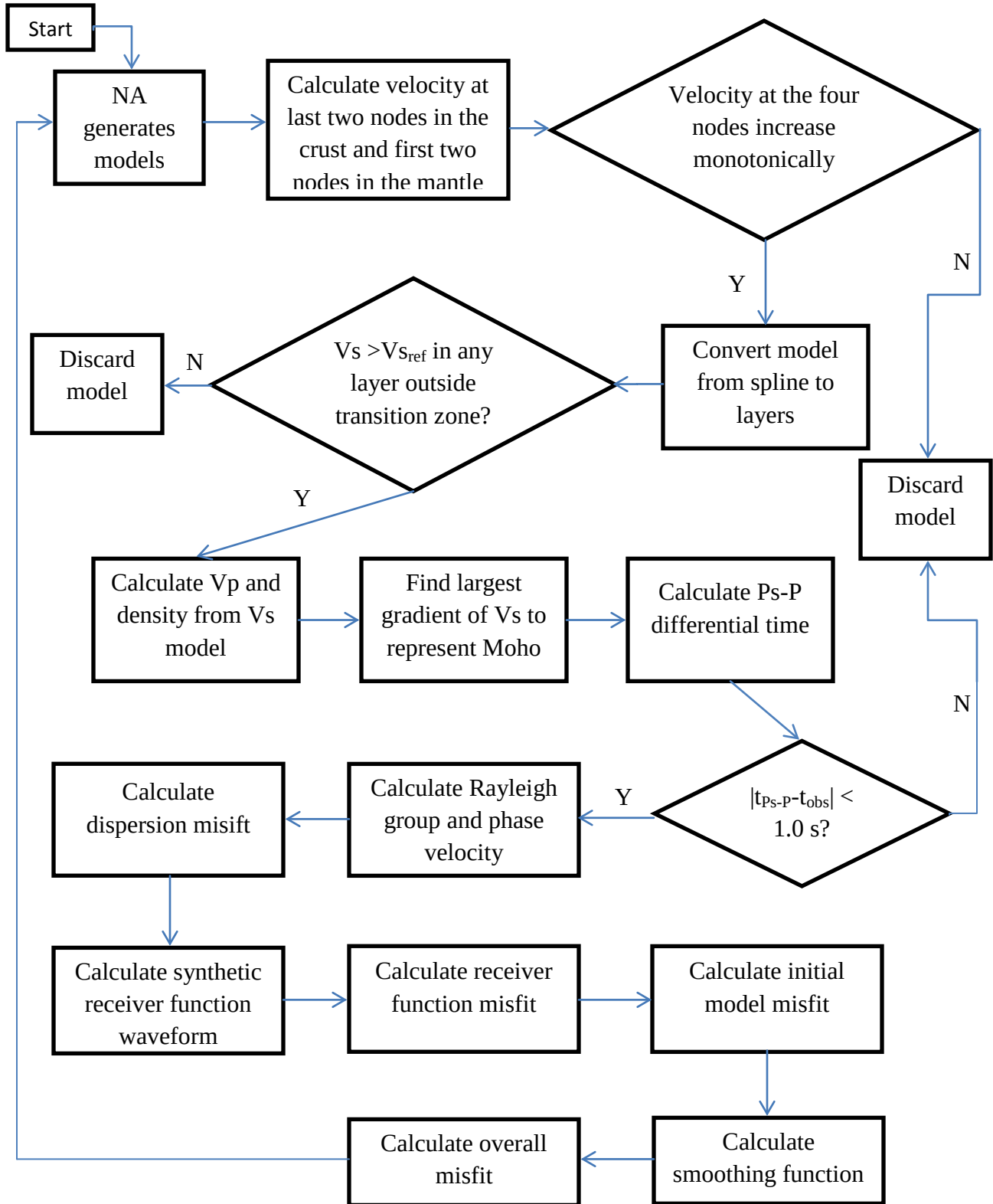


Flow chart of 3D joint inversion scheme, continued



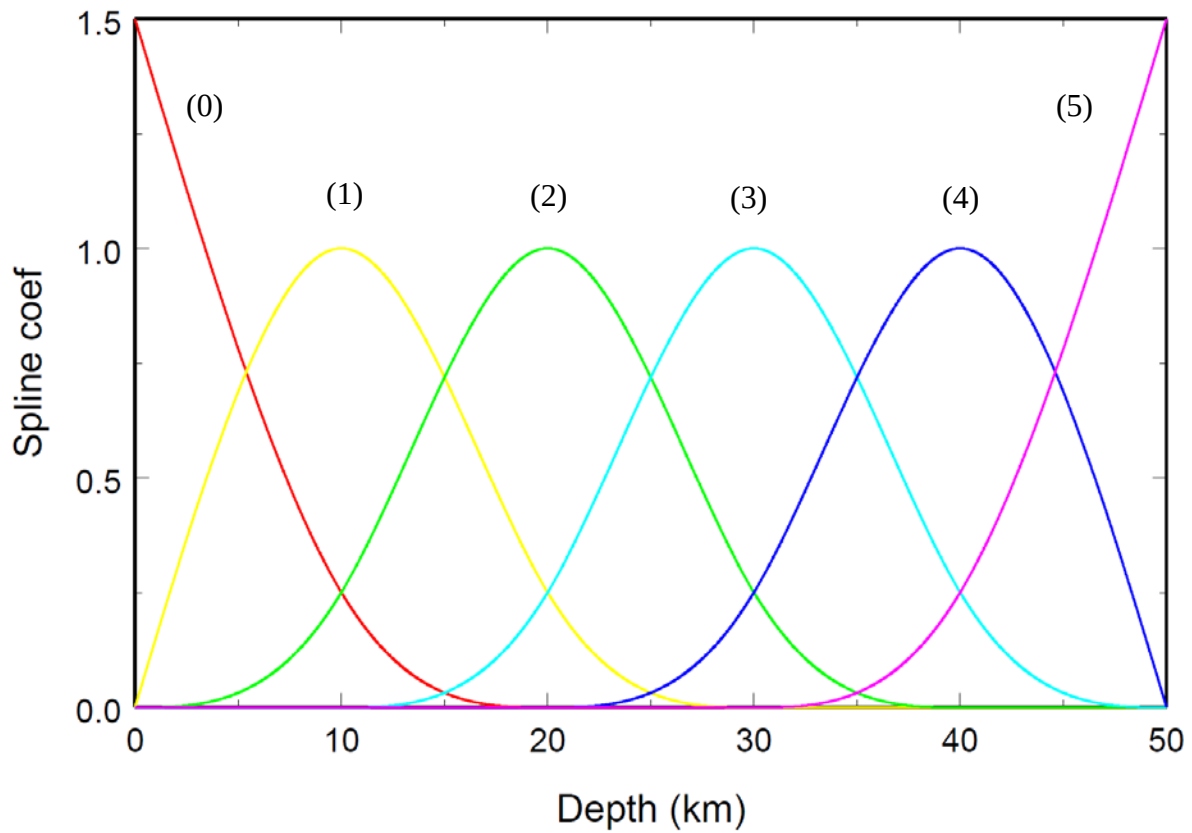
APPENDIX B: SUPPLEMENTARY TO CHAPTER 3

B.1 Flow chart of joint inversion using dispersion and receiver function data

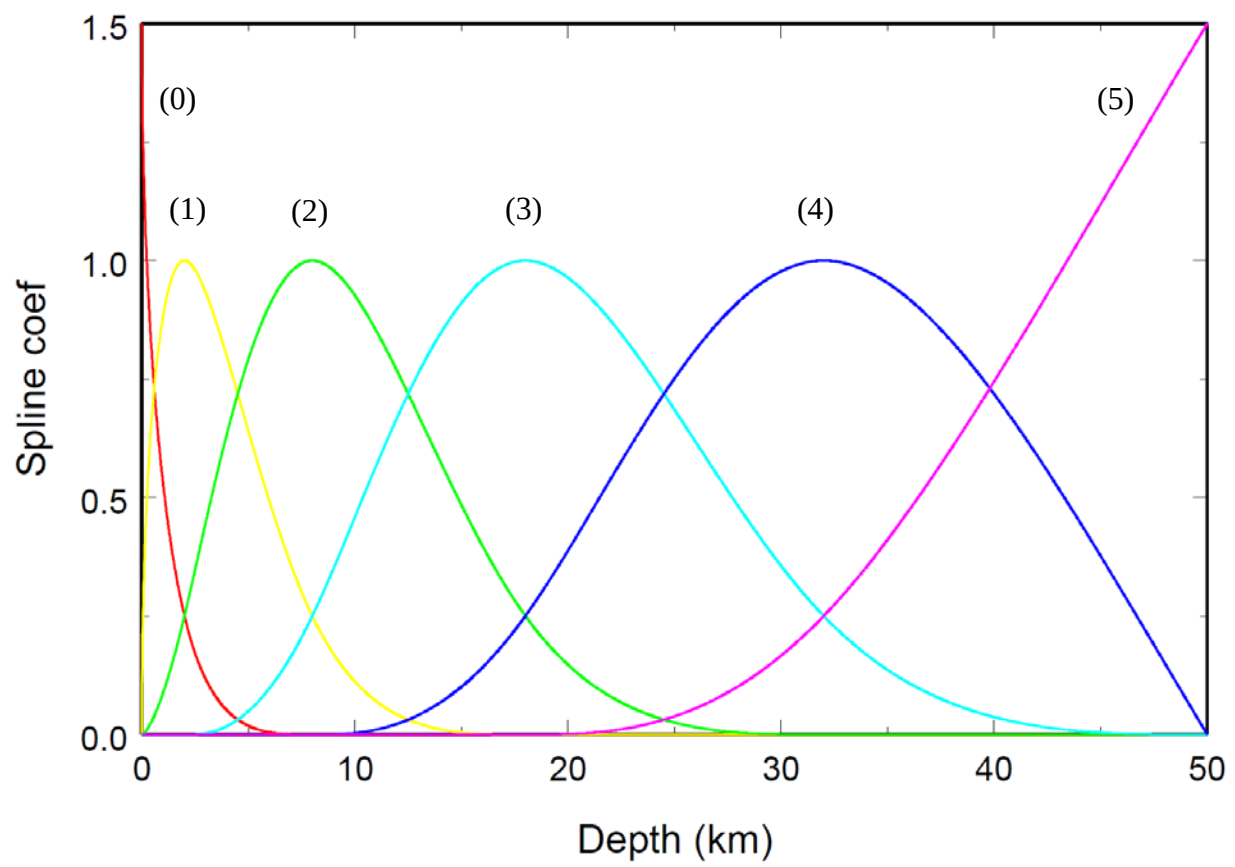


B.2 Cubic spline basis functions

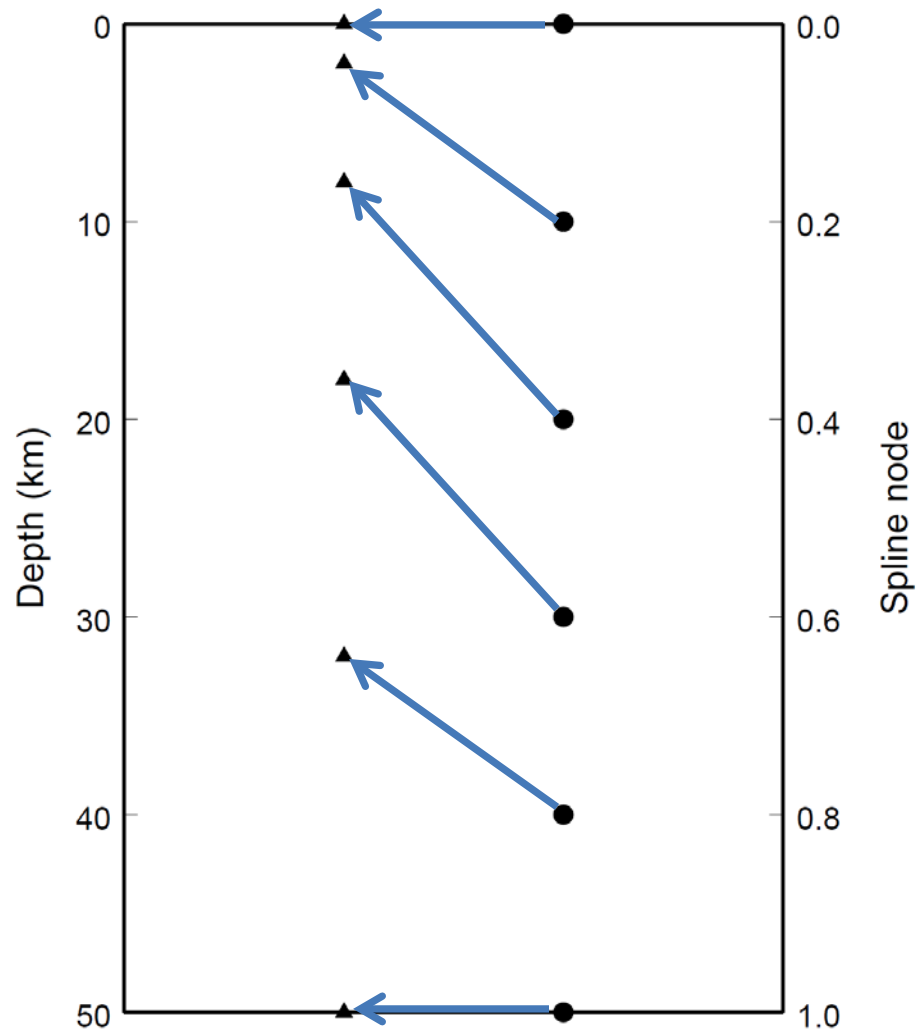
Examples show the spline basis function. Six spline nodes are assumed over a depth range of 50 km. (a) shows spline basis functions of traditional evenly distributed spline nodes. Different colors denote different node. (b) shows spline basis functions of stretched spline nodes with stretching parameter of 2. (c) shows the mapping from spline domain between 0 and 1 to depth domain from 0 to 50 km. Arrows indicates the one on one node mapping.



(a)



(b)



(c)

B.3 Parallel computation

Parallel process using message queue

Message queue is commonly used for inter-thread or inter-process communication. It uses a queue to pass control or content. The structure of message queue involves initialization, dividing tasks and sending messages, receiving messages and executing tasks, and finish. In the message queue, the sender and the receiver of a message do not need to interact with the message queue simultaneously. Messages are stored in the queue until the receivers retrieve them. Therefore the message queue is best suitable for asynchronous messaging. Once all the messages are retrieved and tasks are executed, the whole process finishes. For the same reason, the task must be self-contained, meaning that everything is done within the task. Message queue can only be used within the processors of the same computer node. In ambient noise correlation, data for every day or every station need to be processed individually, thus it is ideal and easy to implement message queue in data processing. In the data filtering and data whitening step, the station list is divided into small groups and sent to different processors (for maximum performance, one processor only executes one process). In the cross-correlation step, station pairs are broken down and evenly distributed to the processors. However, the loop of calendar days for each station is performed within the assigned processor. Experiment suggests the computation for cross correlation can speed up by about 6 times when 8 processes are involved.

Parallel process using MPI

Message passing interface (MPI) is a language independent communication protocol for parallel programming by sending and receiving messages. It is typical used for supercomputers where memory is not shared between nodes. Other advantages of MPI are its portability and speed, making MPI the most popular message-passing application for high performance computation (HPC). MPI provides flexible communication functionality between processes and therefore has a wider range of applications than the message queue. MPI is adopted in the Neighborhood Algorithm (NA) to increase computation speed. In NA, the master processor (again, it should be a process strictly speaking) generates a series of samples and distributes the samples evenly to the slave processors for forward calculation and misfit evaluation. Each slave processor passes the misfit value corresponding to each model back to the master processor. The master processor collects all the misfits from different slave processors and regenerates new models for the next iteration based on the previous results. This procedure goes on until all the iterations are finished. MPI significantly reduces the forward modeling time.

APPENDIX C: SUPPLEMENTARY TO CHAPTER 4

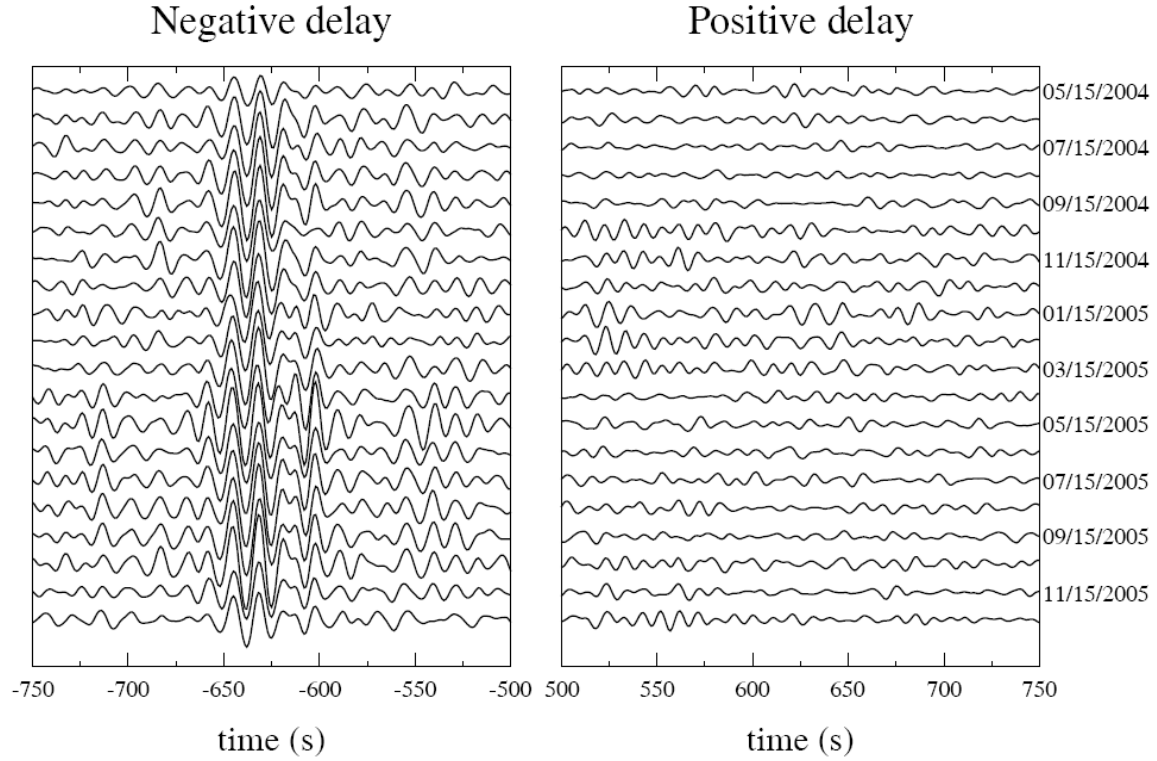


Fig. S1. Asymmetry in Empirical Green functions for the CHTO-PSI pair. Asymmetry indicates noise sources are mostly from the south (Indian Ocean). The similar amplitude of noise in the positive side relative to the signals in the negative side indicates there is no systematic or sudden change of noise sources and that Rayleigh waves retrieved from noise correlation are stable.

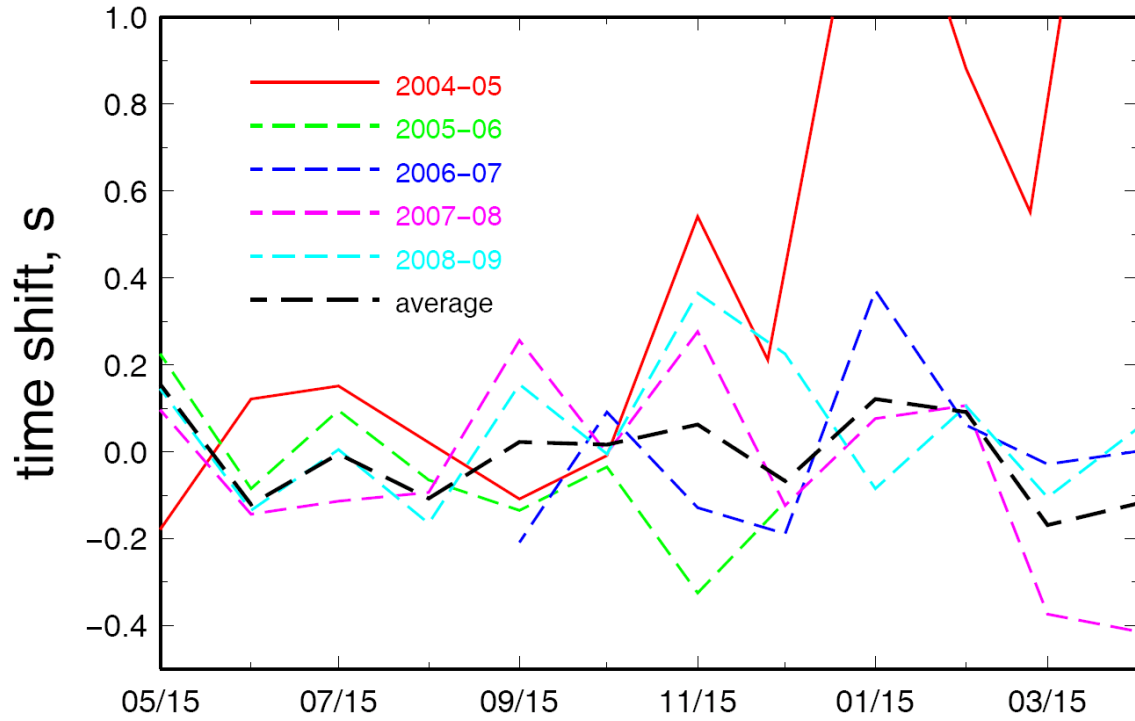


Fig. S2. Time shifts of the CHTO-PSI pair for different calendar months. Shown are time shifts from May through April (other years have been shifted to align with year 2004). In order to compare with the unusual signal in Nov. 2004, the average of the first six months (from May through October) in the corresponding year is removed from the data of that year. Data points are plotted in the middle of each month. The black dashed trace is the average using data from May 2005 through Apr. 2009. Note there is no obvious seasonal variation and Nov. 2004 shows the largest anomaly among all the time periods except the months after 2004 and 2005 event.

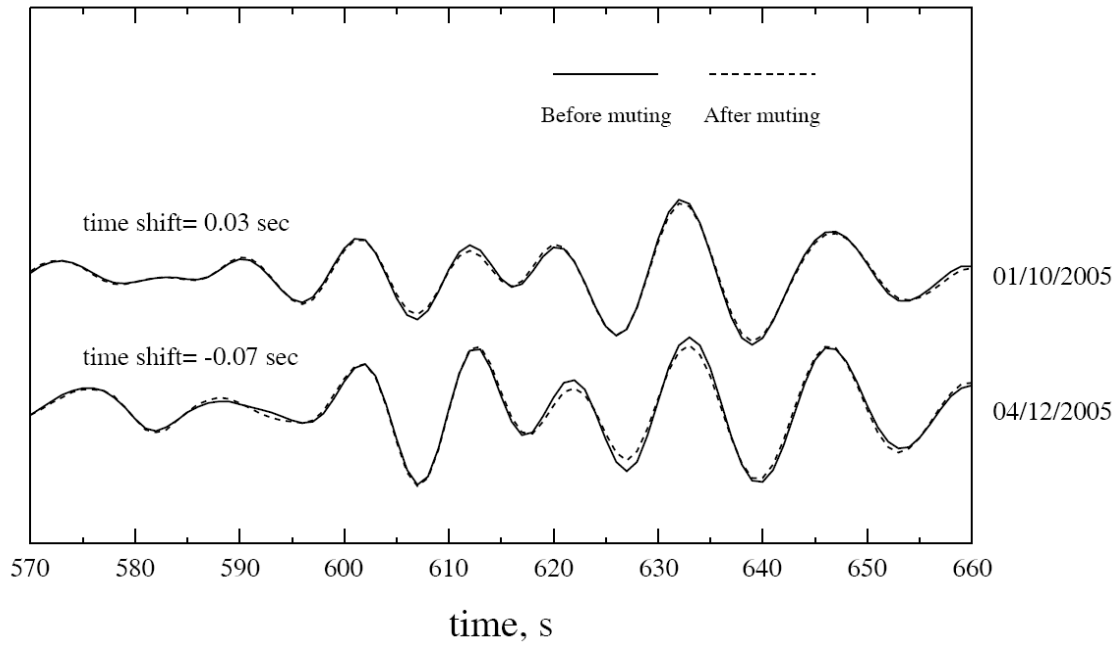
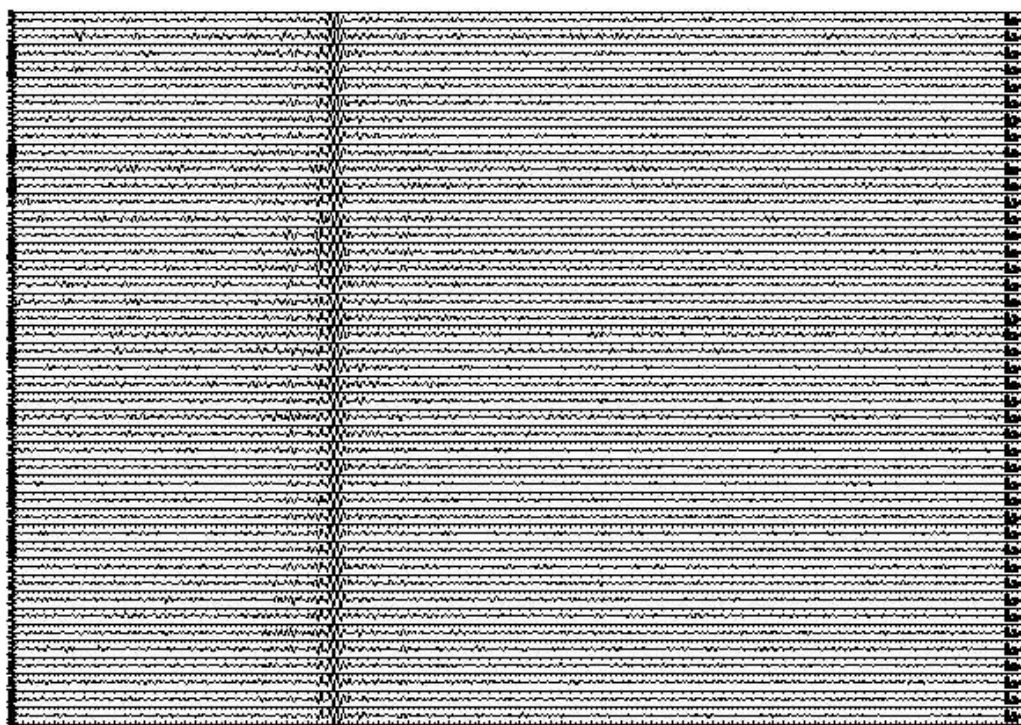
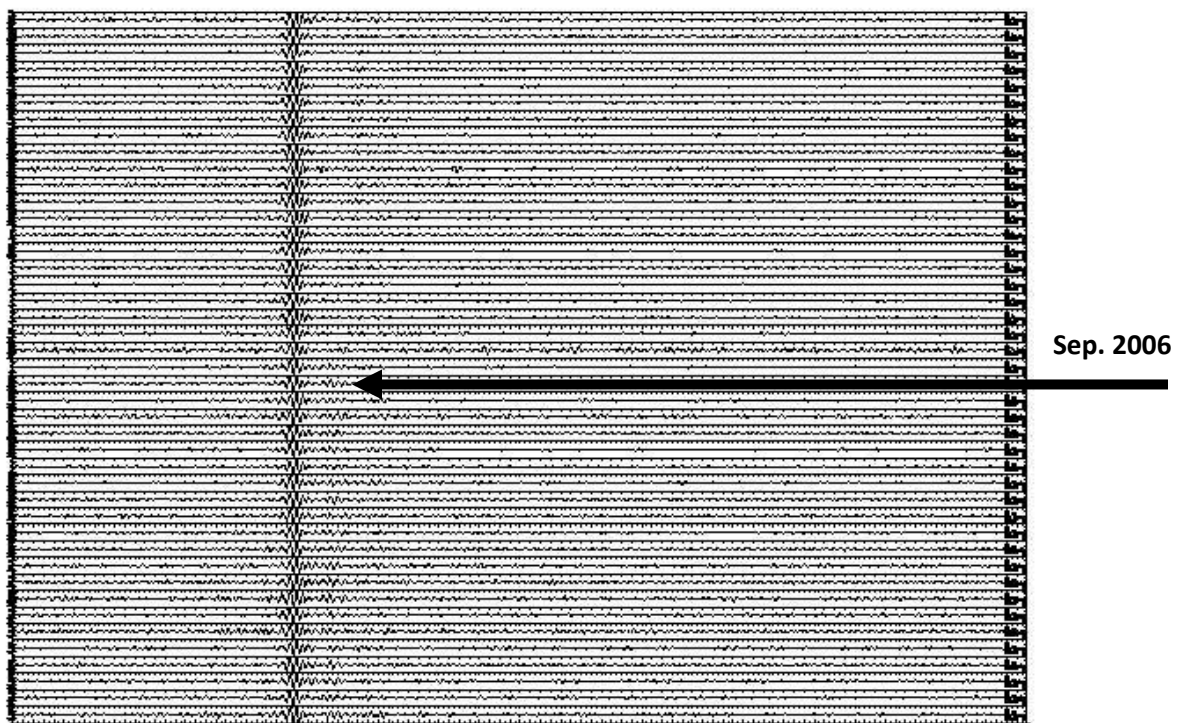


Fig. S3. Comparison of EGFs of the CHTO-PSI pair with and without major aftershocks using one month of data after 2004 and 2005 events, respectively. Note the waveforms before (solid line) and after (dashed line) major aftershocks windowing are almost identical with only slightly change in amplitude. The relative time shifts before and after the muting are 0.03 s for 2004 event and -0.07 s for 2005 event, respectively. The similarity of the waveforms indicates that the time shifts we observed after 2004 and 2005 events are not likely from large aftershocks.



(A)

Fig. S4



(B)

Fig. S4

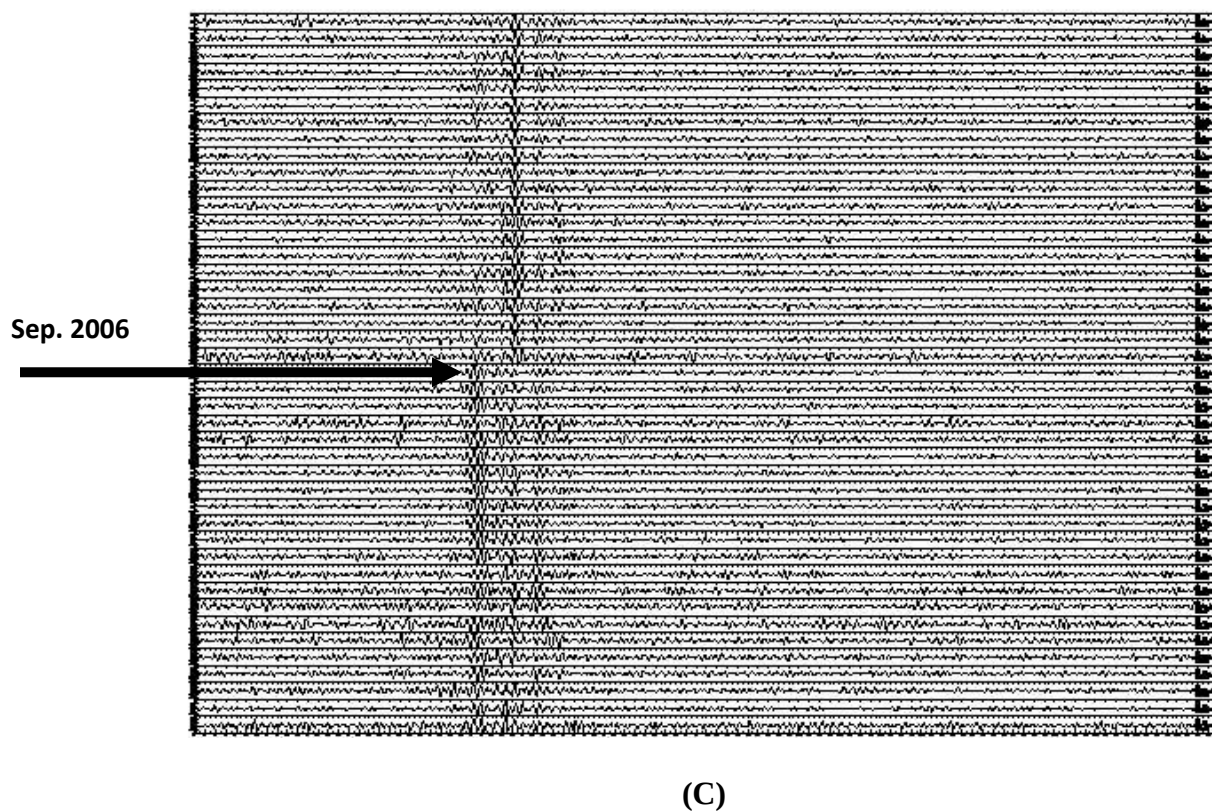


Fig. S4. Symmetric part of EGFs for the CHTO-PSI pair from correlation of Z-Z component (**A**), T-T component (**B**), and R-R component (**C**). Shown are all the data from Apr. 2004 to June 2008. Black arrow marks the data break (no data during Jan.-Aug. 2006). The large time window (0 to 2000 s) is used to show better the change of the overall waveforms. Z-Z and T-T correlations generally are consistent throughout the whole time period (**A** and **B**). Love wave energy leaks into R-R correlation to become the dominant energy after the break (**C**). Note there is also an apparent change in the coda of Love wave after the data break (**B**).

Movie. The movie (in a separate file: 2007_sumatra.wmv) shows spatial and temporal evolution before and after the Sept. 2007 Sumatra earthquake. Each clip shows the distribution of the pattern of the rms time shifts for the two months period that are calculated using the method described in the Fig. 5 caption. The movie is in WMV format.

AUTHOR'S BIOGRAPHY

Zhen Xu graduated from Nanjing University, China in 2003 with a Bachelor of Science in Geosciences. He completed his Master of Science in Geophysics also in Nanjing University, China in 2005 before he joined the Department of Geology in University of Illinois at Urbana-Champaign as a PhD student under the guidance of Prof. Xiaodong Song. He studied and worked in Department of Geology in University of Illinois at Urbana-Champaign as a Research Assistant and a Teaching Assistant for 6 years. He is a student member of AGU, AAPG, and SEG.